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PETROLOCIG RECONNAISSANCE IN THE PRE-CAMBRIAN OF THE WESTERN PART OF THE WIND RIVER MOUNTAINS, WYOMING

BY

CHRISTOFFER OFTEDAHL.[1]

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Abstract. The pre-Cambrian of the west central part of the Wind River Range consists essentially of medium-grained and porphyroblastic, granitic rocks, faintly gneissose, but with no regionally constant strike. Patches of migmatite gneiss and agmatite occur throughout and seem to prevail in the peaks along the Continental Divide. One, possibly two, generations of older basic intrusions, mostly irregular dikes, have been partly granitized. Scattered late pre-Cambrian dikes are unmetamorphosed.

Introduction.

The Wind River Mountain Range, 100 miles long and 30 miles broad, consists mostly of pre-Cambrian rocks. Very little has been previously published about these rocks. The only petrographic descrip-

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[1] The present paper was written while the author was visiting assistant professor of geology at University of Illinois in 1950.

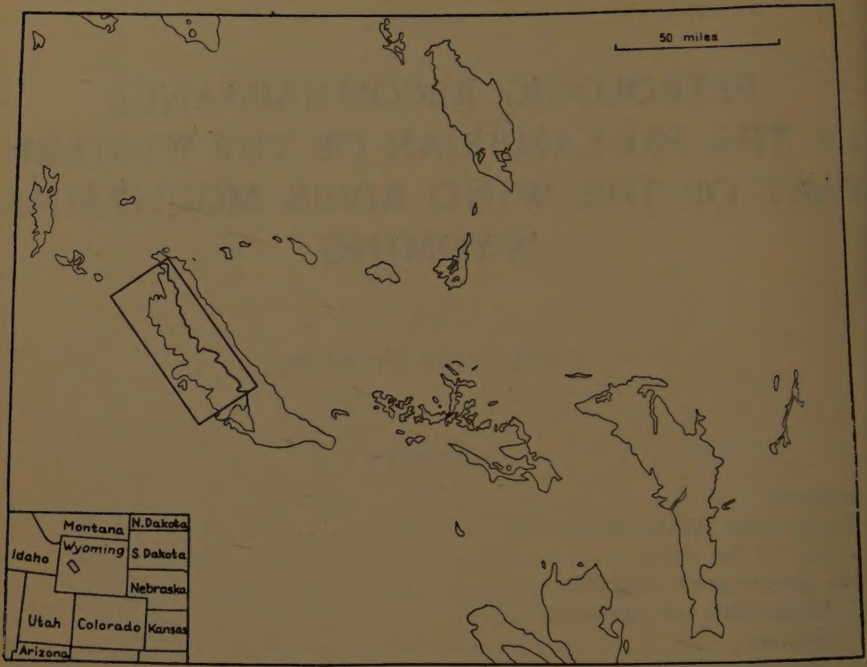

Fig. 1. The state of Wyoming, with the pre-Cambrian areas outlined. The rectangle shows the part of the Wind River Range, shown on Fig. 2.

tion is from the gold district at the southern end of the range, north of Atlantic City. Here Spencer (1916, pp. 14—21) describes granites, metamorphosed sediments and basic intrusions. Two recent publications have some interest for the present description. Richmond (1945) describes the sediments of Cambrian to Tertiary age which border the pre-Cambrian from New Fork Lakes to north of Green River Lakes. Granite and gneiss are distinguished in the adjacent pre-Cambrian, and a great number of shear zones are drawn on the map. Nearly exactly the same features are described in a publication by Baker (1946), who also mentions the existence of grey gneiss, fine-grained to very coarse-grained granites and quartz monzonites.

The present author had the opportunity of doing some field work in the U.S., and chose to undertake a petrologic reconnaissance in the Wind River Range, in order to establish the main petrographic status of the area. This may be of some importance for regional correlation of the many pre-Cambrian areas in the states of Montana and Wyoming and may serve as a base for future detail work.

Acknowledgement.

The field work for the present paper was carried out during 1½ months in the summer of 1950, supported by a grant from the Research Board of the Graduate College, University of Illinois. I am indepted to Prof. T. F. W. Barth and Dr. A. Fredrickson for corrections and suggestions.

Survey of the Rocks.

The rocks of the investigated area fall into three groups:

I. The granite-gneiss complex.
II. Dikes and irregular bodies of meta-gabbro.
III. Late pre-Cambrian, unaltered dikes.

The granite-gneiss complex is very heterogeneous. It contains homogeneous rocks of granitic composition and "soupy" gneisses with continuous transition between dark and light patches and bands (migmatite). Transitions between the two mentioned rock types are most frequent. One typical variety contains sharp-edged fragments in a homogeneous granitic rock (agmatite). There seems to be little possibility for mapping any of these rock types; they always grade into each other by transitions. The same is the case with the porphyroblastic or medium-grained granitic rocks. Within both groups of rocks there are granite, granodiorite and quartz-diorite occurring in a rather irregular way, so that only vague hints about the regional distribution can be given.

Dikes and irregular bodies of meta-gabbro occur all over the investigated area. The rocks are usually basic amphibole diorites, with more or less pyroxene, often as cores in the amphibole individuals. Consequently the rocks are supposed to be altered gabbro, here called meta-gabbro. They are always metamorphosed and to some extent granitized. On account of varying degrees of granitization there might be two generations of intrusions, both of which are supposed to be younger than the general metamorphism and meta-

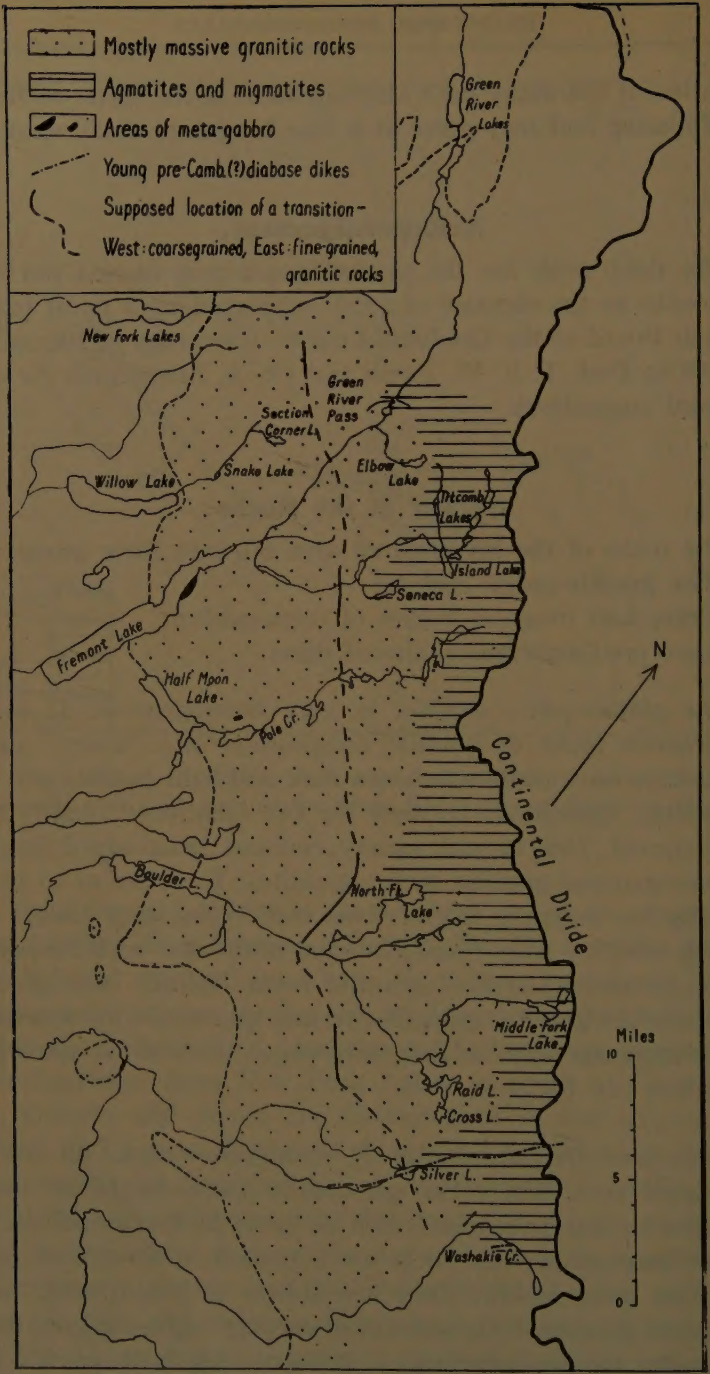

Fig. 2. Geological sketch map of the central portion of the Wind River Range, southwest of the Continental Divide. The blank area to the left is glacial deposits and mostly Tertiary sediments.

somatism in the area. These intrusions may be related to the intrusions described by Spencer (1916, pp. 19—21) from the southern end of the range.

Late pre-Cambrian, unaltered dikes, have been found in a few places in the interior part of the range. One of them is rather straight for eight miles, the two others have a limited length.

In the following these three rock complexes are described petrographically, and the main distribution of the various rock types is outlined.

The granite-gneiss complex.

The various rock types within this complex show transitions from the one to the other, and therefore a description of the rock variation can only be given in general terms.

There is a certain tendency towards a zonal arrangement of the crok types in the investigated area. The high peaks along the Continental Divide consist to a large extent of migmatites and agmatites. The mountainous zone above and just below the timberline and southwest of the high peaks is made up of a mixture of medium-grained granitic to dioritic rocks with areas of migmatites and agmatites. In the southwestern marginal zone (central and lower part of the belt, 7000′—9000′) more homogeneous granitic rocks prevail, although there are many small areas with migmatites and scattered inclusions. Porphyroblastic rocks also occur essentially in the marginal zone.

In the following the rock units are described as to their macroscopic and microscopic appearance and occurence: Migmatites and agmatites, diorite, medium-grained and porphyroblastic granitic rocks.

Migmatites and Agmatites.

The migmatite gneisses have a highly varied appearance. One type contains dark, elongated fragments of an older rock floating in lighter gneissic rock. The fragments have diffuse contacts, and sometimes they appear only as dark shadows. Another type has undergone extensive folding, so that it consists of alternating dark and light stripes, swirling around in complex patterns. All transitions between these two types exist. Generally there is no strike of the gneiss

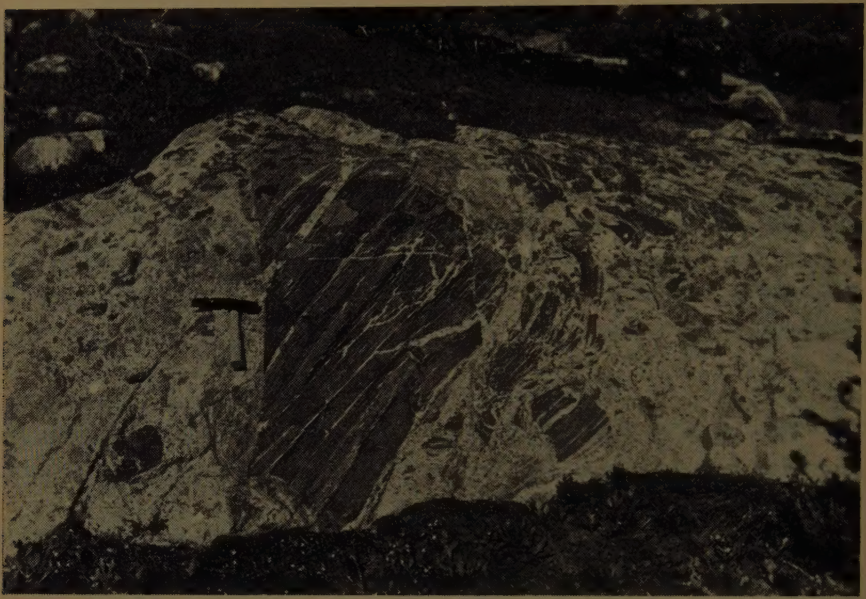

Fig. 3. Typical agmatite-migmatite gneiss, just north of Middle Fork Lake. The central, big fragment is only invaded along its schistosity planes by aplitic veins. A "soupy" gneiss flows around the more assimilated fragments to the right and left.

foliation that persists over as much as 100 feet, but rarely one strike direction may prevail for a mile or two.

The first type seems to prevail in the area from Palmer to Elbow Lake, while the highly folded type is beautifully exposed northeast of Island Lake towards Fremont Peak.

Some places (e.g. at Middle Fork Lake) the inclusions have partly a sharp contact, and thus grade towards the agmatites. True agmatites, with angular fragments having sharp contacts, occur only as patches or outcrops in the migmatite area. The best example is from the mountains just north of Green River Pass, where huge blocks which have fallen down into the valley show that whole mountain sides may consist of agmatite.

The mineralogical composition has been estimated in a number of thin sections, giving a rough, but sufficient basis for classification.

The basic inclusions are characterized by their content of augite, which is commonly partially altered to hornblende, thus suggesting that they are meta-gabbros. The hazy dark patches or stripes have a composition from dioritic via quartz-dioritic to grano-dioritic. White aplite — pegmatite veins invade the inclusions in the agmatites. These veins vary in composition from pure quartz diorite to ordinary granite.

A good example of the process of assimilation of the fragments is obtained from the northern shore of Middle Fork Lake. Here dark fragments of meta-gabbro with diffuse contacts are imbedded in a migmatite gneiss with a patchy appearance due to lighter and slightly darker areas. The lighter colored rock is grano-dioritic whereas the darker rock has a dioritic composition with big hornblende crystalloblasts, some with pyroxene core. But the white aplite veins cutting the fragments are granitic with only 20 per cent plagioclase. These facts, together with the observation that the fragments show all gradations from angular black inclusions to barely visible shadows strongly support the view that this process of assimilation is of the replacement type, that is, granitization.

It is a question what other kinds of rocks have been assimilated, but it seems likely that the migmatite gneisses represent granitized sediments. Only one occurrence of true sediment, partly granitized, has been found. In the trail at the southwestern end of Elbow Lake, a light gneissose rock forms a dike-like body in the diffuse migmatite gneisses. The dikelike body has a pronounced foliation which is concordant with that of the gneiss. Under the microscope the light rock consists of clastic quartz grains in a fine-grained groundmass, containing epidote and sericite. Some 30—50 per cent microcline, absolutely clear, is obviously formed by the metamorphism or by metasomatism. Thus the lighter appearing rock is no doubt a meta-quartzite. From the Green River Lake area, Baker (1946, p. 567) reports: ''Schists, dominantly amphibolitic, and green quartzites are rare.'' There is good reason to believe that at least part of the gneisses consist of granitized sediments, where very little of the primary structures are left.

Diorites.

Dioritic rocks are rare in the described area. The only occurrence of what seems to be a larger mass is situated around Raid Lake and some miles to the southeast and northeast. The rock is an ''igneous-looking'', hornblende-bearing diorite. It is essentially massive but contains rare schlieren of nearly digested inclusions. Patchy migmatite structures are also seen. The diorite is characterized by a bluish green metamorphic hornblende.

Rocks of dioritic composition (in the sense of Johannsen) have been found in a few places, both in the migmatite gneisses and in the medium-grained granitic rocks, and many rocks from both of these groups are rather close to a diorite in composition, containing 5 to 10 per cent quartz or 10 to 20 per cent microcline or both.

Medium-grained granitic rocks.

Along the southwestern margin of the pre-Cambrian area granitic rocks prevail. They are usually massive, but have sometimes a faint schistosity. Nearly complete absence of migmatite structures is characteristic. The rocks range in composition from microcline granite to quartz diorite. The textural varieties range from medium-grained rocks to porphyrblastic rocks with big feldspar idioblasts. Intermediate textures are rather rare, and the two varieties may be described separately.

The medium-grained granitic rocks have a varied color, depending on the distribution of the femics. The most interesting microscopic features are the relations between the feldspars and the types of femic constituents.

The relations between the two feldspars, microcline and oligoclase (sometimes sodic andesine), is considered highly indicative of the origin of the rocks. Thirty samples of medium grained granitic rocks were sectioned. Fourteen of them are inconclusive, showing a eugranitic texture which may be either metamorphic or igneous. In some sections both feldspars are clear, in others both are hazy due to incipient alteration, e.g. by formation of tiny epidote and sericite grains in oligoclase and microcline, respectively. The remaining 16 sections show indications of replacement phenomena. They may be grouped as follows:

Five sections contain water-clear microcline and altered plagioclase individuals. The alteration is shown by dark clouds of either iron ore or faint saussuritization. This suggests that the microcline was formed after the formation of the plagioclase, or after a metamorphism of it.

Four sections show clearly that microcline is replacing plagioclase.

The microcline grains are irregular and send armlike off-shoots into the anhedral plagioclase grains. At the same time square-shaped microcline patches form within the plagioclase tablets.

In eight sections myrmekite occurs in the border zone of plagioclase individuals, where these are in contact with microcline grains (the pre-microcline myrmekite type of Drescher-Kaden, 1948, p. 39). There is a rather general agreement that myrmekite is typical for the metasomatic feldspar replacement (see Drescher-Kaden, 1948, p. 14—39), although myrmekite may form from late magmatic solutions in igneous rocks.

As to dark minerals the rocks contain only biotite (12 sections), biotite and hornblende (3), only hornblende (2), or only chlorite (4); and four sections show grains of augite more or less completely altered to hornblende and biotite.

The granitic, medium-grained rocks show that at least parts of the investigated area have undergone an extensive potash metasomatism. The primary rocks possibly were pyroxene quartz-diorites which have been subjected to a retrograde metamorphism, or, more likely, they could be sediments extensively granitized.

Porphyroblastic granitic rocks.

The porphyroblastic rocks have usually a faint gneissic structure and contain large feldspar tablets of rectangular shape and frequently a length of one to two inches. In places the shape is irregular or more eye-like, but typical augen gneisses are rather rare.

These rocks are called porphyroblastic because the big feldspar crystals show the same features as described above, essentially replacement texture and myrmekite, that is, they have grown by metasomatism. The crystals are composed of microcline which is slightly microperthitic in most sections. A few sections show plagioclase idioblasts, and there are a few cases where such crystals are being replaced by microcline.

It is interesting to note that in most cases there has been a potash metasomatism, whereas some rocks suggest that there has been an earlier soda metasomatism. The formation of the porphyroblastic rocks took place either by metasomatism of the metamorphic sediments and basic igneous rocks, or by a recrystallization of the igneous granodiorites. Reliable evidence on this point is not ascertain-

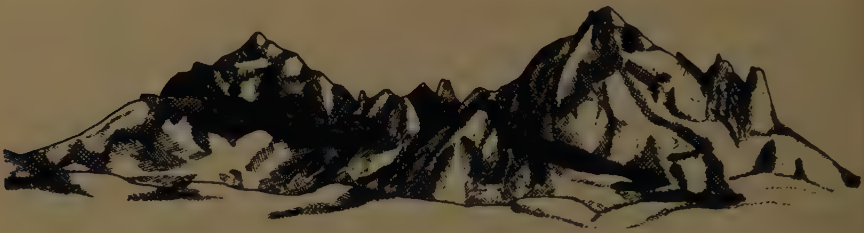

Fig. 4. Big dike in the northern slopes of Woodrow Wilson Peak, situated half way between Island Lake and Elbow Lake.

able from thin section studies but only obtained from field observations such as porphyroblastic rocks fading into augengneiss and the granitization of the large basic inclusions, to be described below.

Basic dikes and irregular bodies.

All over the investigated area bodies of dark, metamorphosed rocks occur in gneissic or granitic rocks. The shapes of the bodies vary between more or less clearly recognizable dikes and sills to huge irregular bodies which may be between one hundred and several thousand feet long. Both dikes and irregular bodies are frequent in the high mountains close to the Continental Divide but rather rare in the marginal zone of granitic rocks.

Dikes. The dikes or dike remnants form dark bodies in the lighter rock, with a width of one to six feet, their ends fading out in the bedrock. Some dikes have been followed for half a mile, but their length is usual between fifty and two hundred feet. They are most frequent in the migmatites of the high peaks east of the High Line Trail between Elbow Lake and Island Lake. Here they have often the character of nearly horizontal sills. In one mountain (see Fig. 4) a big dike traverses the whole mountain side, with a probable dip of 40° to the west. In the mountains south of Island Lake many sills are seen, and in one place (west slope of Mt. Lester)[1] four parallel sills are seen at various heights. In the intermediate zone they are less frequent, and in the marginal, granitic zone rather rare.

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[1] Some locality names in brackets are not on sketch map of Fig. 2; the reader is referred to the Forest Service maps or the Fremont Peak and the Mt. Bonneville quadrangles.

Generally speaking the dikes are not post-tectonic; they are all metamorphosed. Their microscopic textures are those of metamorphic rocks, with a composition mostly like basic diorite. Plagioclase (35—45 An), and augite more or less altered to amphibole are the main constituents. The observations of the appearance in the field give the impression that there are two generations of dikes. The first generation is metamorphosed and considerably altered by crystallization of metasomatic quartz and feldspar. The second generation is metamorphosed only, or granitized in a few places.

The difference between the two generations is demonstrated in one outcrop (at the trail to Island Lake, north of the third lake above Seneca Lake). Here a massive meta-diabase, intruded by pegmatite dikes, cuts a dike which is schistose and is heavily invaded by thin pegmatite veins. This shows that the old dike is metamorphosed, deformed, and metasomatically altered, whereas the younger is only metamorphosed. A clear age difference is also found between two basic dikes running parallel at the trail to Bald Mt. Basin (just east of Pole Creek). A fine-grained dike contains fragments of the neighboring coarse-grained dike, and the latter has been subjected to a potash metasomatism, with the formation of scattered white streaks and spots of microcline, quarter to half an inch long, and some 10 per cent microcline as small grains, seen in thin section. Again only the older dike has been metasomatically altered.

It is not easy to classify all dikes observed as either first or second generation dikes. The main impression is that most dikes, also the above-mentioned sills, belong to the second generation. But the partial assimilation of the first and the second generation dikes is characteristically different and is described below from two occurrences.

North of Cross Lake a black dike crosses the High Line Trail. The central portion consists of a nearly medium-grained altered diabase with clear andesine plagioclase and augite which is mostly altered to green hornblende. In the border zone the alteration has has gone further. The pyroxene has disappeared and some chlorite has formed, and it also partially replaces plagioclase. The dike is cut by a very coarse pegmatite vein and by a few faults with a horizontal displacement of up to two feet. The contacts to the bedrock are sharp and straight-lined, except in a few places (see Fig. 5). In one occurrence a vein of the bedrock invades the dike with diffuse

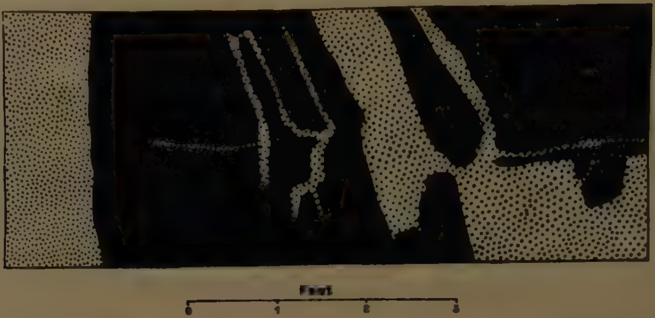

Fig. 5. Outcrop where the bedrock (left, dotted) invades a basic dike of second generation (black). The contact between the two rocks is fairly sharp. Location 100' south of the High Line Trail, north of Cross Lake.

contacts. Another outcrop shows a gradational transition over two feet. The best evidence for activation of the bedrock gneiss after the intrusion of the dike, is obtained from the observation of angular fragments of the dike floating in the gneiss.

One dike is conveniently located at the trail on the northern shore of Half Moon Lake, 600 feet east of the end of the road. The dike, some 5 feet wide, crops out over a length of about 100 feet. The rock looks like a melaphyre, with scattered black crystals, 2—4 mm. across, in a dark ground-mass. The contact looks mostly sharp and magmatic, but some parts of the contact show even gradations, other parts show a 2 inch intermediate zone. Under the microscope the black phenocrysts appear to consist of green hornblende.

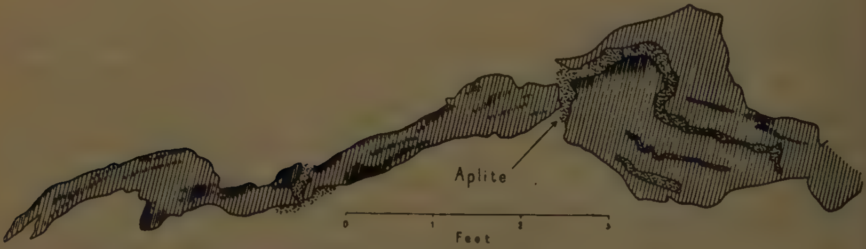

Fig. 6. First generation dike, at the trail south on Doubletop Mountain, between Willow Lakes and Section Corner Lake. The dike is strongly folded and is cut and invaded by aplite. Dark areas are little altered, and the lighter areas are metasomatically altered to a dioritic composition.

In the ground-mass most pyroxene grains are altered to hornblende. The plagioclase is to a considerable extent replaced by quartz, possibly as much as 30 per cent. The granite, taken one inch from what seemed a fairly sharp contact, proved to have big crystals of hornblende, the same pseudomorphs after pyroxene as in the dike itself. These facts indicate a migration of basic matter from the dike into the bedrock, and probably also an invasion of quartz and feldspar into the dike from the adjacent granite.

Irregular basic bodies. Big bodies of metamorphosed, basic rocks are scattered all over the region. They seem to be especially big in the high peaks northwest of Freemont Peak up to Green River Pass. When invaded by aplite veins, they form the agmatites, mentioned on p. 6. All samples of such rocks, even from dark inclusions only a foot across, contain some pyroxene, with hornblende as an alteration product of pyroxene. Thus it is obvious that all dark inclusions are partially metamorphosed igneous rocks, mostly gabbros. It is then near at hand to suppose that the migmatite gneisses all are derived from granitization of a gabbro complex. That possibility seems less likely, though, because the small meta-gabbro lumps, 30—50 feet across, show approximately the same degree of metasomatism as the dikes. They are altered to feldspar porphyroblast schists only in the border zone, or small lumps 10—20 feet across may be completely changed into such rocks. Other bodies seem only to have been metamorphosed. The author's interpretation is that the basic intrusions are later than the general metamorphism and metasomatism, and that there are two generations of intrusions, corresponding to two generations of dikes.

One body of meta-gabbro situated at the eastern shore of Fremont Lake in the middle of the northern half of the lake demonstrates granitization particularly well. The length along the shore-line is 1½ mile. A sample of what seemed to be only metamorphosed gabbro, consists of andesine (40 An), 50 per cent biotite, and 10 per cent quartz, obviously replacing the plagioclase. In another greenish meta-gabbro the femics are completely altered to chlorite, which also replaces the plagioclase to a considerable extent. Replacement quartz makes up 15—20 per cent of the rock, still without any formation of microcline.

The more ordinary granitization is demonstrated by the northern-

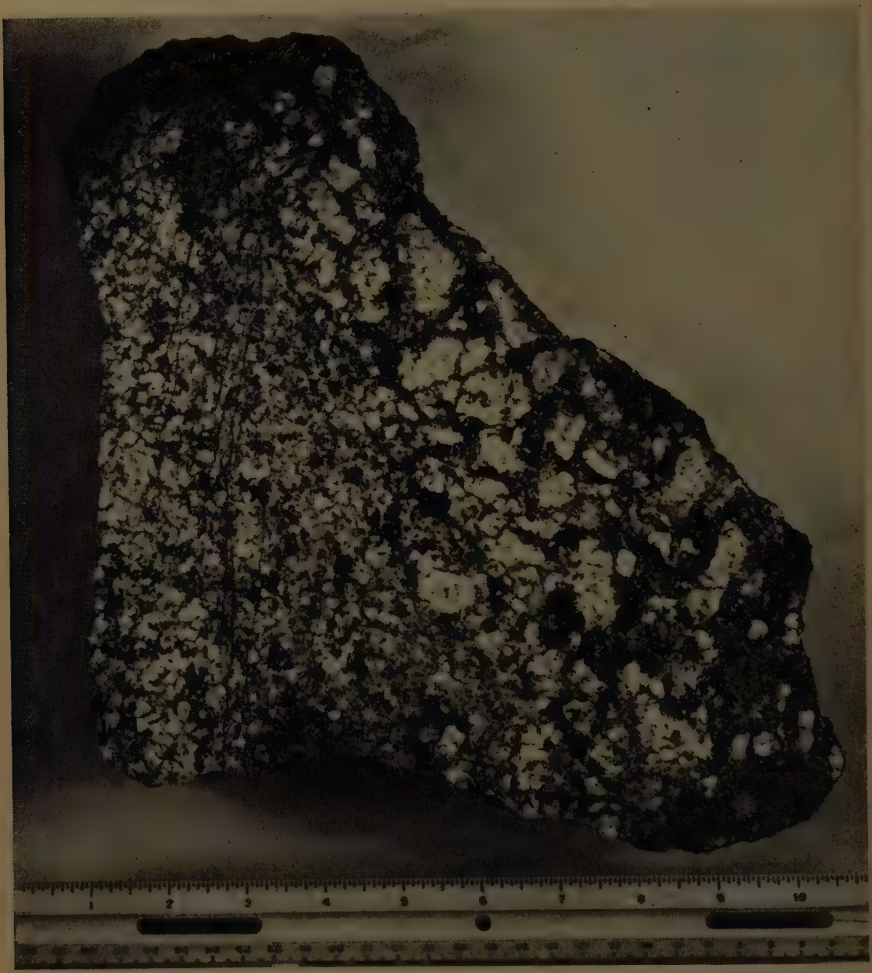

Fig. 7. Specimen from northern part of the meta-gabbro at Fremont Lake. The lower right tip is a greenish dioritic rock; towards the left microcline porphyroblasts and quartz grains (not seen on photo) start growing, so that the central left part of the specimen is a microcline granite with "replacement appearance".

most part of the massif, where rectangular microcline tablets start growing, increase in number, and, being accompanied by quartz, change the dark rock into a porphyroblast granite. The latest thing to happen was the invasion of microcline granite aplite.

Late pre-Cambrian diabase dikes.

Three dikes have been found which cut through all other structures. "The Black Dike", which crosses Silver Lake, can be seen to continue at least 8 miles; it has a beautiful ophitic structure, and it is a typical diabase. The easternmost dike at Washakie Creek is rather unusual, with essentially monoclinic and orthorhombic pyroxenes in a glass base.

The existence of a few late pre-Cambrian diabase dikes is mentioned by Baker (1946, p. 568) from the Green River Lake area and two dikes have been mapped by Richmond (1945); the northern one is situated at the northern boundary of the map (Fig. 2) and the other occurs southwest of Green River Lakes and is covered by the legend in Fig. 2.

Discussion of the origin of the rocks.

The observations leave no doubt that a regional metamorphism and metasomatism has taken place in the region described. By metasomatic replacement geosynclinal sediments were altered to granodiorites, possibly part of the rocks were derived from granitization of gabbros.

The last type of metasomatism was a K-replacement, with formation of microcline at the expense of oligoclase. But the existence of granodiorites with big plagioclase porphyroblasts seems to suggest an earlier Na-metasomatism. However, the irregularly occurring K-metasomatism gave birth to a region of granitic rocks with rapid variation from granites to granodiorites, and quartz diorites, and even diorites. The transition between these rock types is usually gradational, but relatively sharp contacts may occur. Thus Baker (1946, p. 567) writes: "With an eruptive contact west of the metamorphism is an extensive granite intrusion overlapped further west by Cambrian". The writer has also seen such an "eruptive contact", where a geologist according to the old classical views sees a porphyritic granite sending apophyses into a medium-grained granodiorite. The contact is not magmatic, however, but metamorphic. A non-believer in replacement processes might interpret such a contact as a metamorphosed igneous contact, but the present author is certain of the following interpretation: During shear movement within one zone of the gneissose granodiorite invading solutions deposited a band of microcline porphyroblasts, changing the primary rock to a porphyroblastic granite with a seemingly sharp contact to the granodiorite.

The type of granitization in the area is not syn-tectonic, since no regional strike developed, but it is obvious that the main granitization took place in the end of the tectonic period, for the nebulitic gneisses usually show folding, which, in places is quite intense. The first generation of basic dikes and bodies belongs to this late tectonic period, whereas the partial assimilation of basic rocks of the second generation seems to belong to a later period of post-tectonic or static granitization.

Generally speaking the rocks of the investigated area are characterized by retrograde metamorphism. Thus the basic bodies show all degrees of alterations of the pyroxene, from equal amounts of pyroxene and hornblende to only a few remnants of pyroxene, the rest being altered to hornblende, biotite, and sometimes also chlorite.

Also the medium-grained granitic rocks show that a similar lack of equilibrium is fairly frequent, mostly with the transition hornblende-biotite, but also with pyroxene-hornblende-biotite, and pyroxene-biotite. But many rocks of this group may have reached equilibrium, carrying only biotite or, in few cases, only chlorite.

The granitic rocks with porphyroblastic feldspars are strikingly different. They all contain a brown biotite as the only femic mineral. This suggests that equilibrium was more easily obtained during the metasomatism resulting in big feldspar porphyroblasts, than during other types of metasomatism or during metamorphism only.

The establishment of two generations of basic intrusions is fairly well ascertained by macroscopic observations, but it is astonishing to find that no difference in degree of metamorphism is found between them. Thus one phase of the retrograde metamorphism must be relatively early, but such metamorphism continued throughout the intrusive periods in a rather irregular way, as also might be expected by static granitization.

In conclusion the geological events may be summarized as follows:

1. An area of geosynclinal sediments underwent late tectonic granitization. The last stage was the formation of aplite — pegmatite veins.
2. Intrusions of basic dikes and irregular bodies occurred throughout the area.

3. The basic intrusions were metamorphosed and partially granitized.
4. A second generation of basic intrusions invaded the area.
5. The younger intrusions were altered by a faint regional metamorphism which was accompanied by replacement processes to a very limited extent. The metamorphism faded out by formation of aplite — pegmatite veins, an inch to a foot wide, the thinner ones irregular and some of the widest ones straight-lined and persistent for miles.
6. The area was invaded by scattered ultrabasic to basic dikes in much later pre-Cambrian time.
7. In Tertiary (?) time the area was uplifted by steep-angled upthrust faults, in places with overthrust character.

The above strict age relations may be modified in several ways. The metamorphism of 3 may be a continuation from 1. In the same way the basic intrusions may be scattered over a time interval during which the metasomatic replacement processes were dying out, thus accounting for what seems to be rocks transitional between first and second generation intrusions.

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PRELIMINARY NOTE ON THE GEOLOGY OF WESTERN DRONNING MAUD LAND

BY

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With 1 map.

Abstract. A series of northeast-trending mountain ranges, with a rock-surface relief of more than 2500 metres, projects through an ice mantle that is in places 1500 metres thick to form the rock massifs and nunataks of western Dronning Maud Land. The oldest rocks found in these ranges belong to a metamorphic complex that probably underlies much of this part of Antarctica. Unconformably overlying the metamorphic complex is an assemblage of flat-lying, clastic sediments; this in turn locally includes, or is overlain by, volcanic rocks. The sediments have been intimately invaded by large sills and dykes of diorite and gabbro. The youngest known rocks are small basaltdykes.

Since the metamorphism of the basement complex, the area has been essentially stable, and has suffered only minor, vertical movements.

Introduction.

The field parties of the Norwegian-British-Swedish Antarctic Expedition 1949—52 visited the outcrop areas of western Dronning Maud Land between Long. 02° E. and 12° W. of Greenwich, and Lat. 71° 08′ S. and 73° 40′ S. Within this region the main rock exposures are confined to an area of approximately 60,000 square kilometres lying east of Long. 09° W. Here a series of nunatak-studded snow ridges, low escarpments, and small dissected mountain ranges are aligned in a general northeast-southwest direction; the mountainous terrain extends beyond the area investigated. This area of inland mountains was discovered and first photographed by the Deutsche Antarktische Expedition 1938—39; and the Norwegian forms of some of the names given by that expedition to prominent topographical features are used in the present note.

Physiography.

On the basis of its topography, the area investigated may be divided into three main units. Northwest of an irregular line running south-southwest from near the nunataks Boreas and Passat (71° 18′ S., 04° 00′ W.), the country is almost completely covered by ice, which rises in a series of gentle, northeast-trending undulations from the floating ice shelf along the coast to a maximum altitude of about 1500 metres. There is very little exposed rock in this region. The few nunataks, all on the crests of the snow ridges, are rarely more than 1 square kilometre in area, and with the exception of Kraulfjella, none project more than 150 metres above the surrounding snow surface. Seismic soundings in this area, however, show the ice mantle to be as much as 1000 metres thick; and the present-day nunataks are seen as merely the exposed points of a rugged rock-surface topography of considerable relief.

To the east of this large area of very sparsely-scattered nunataks, between Lat. 71° 30′ S. and 73° 00′ S., lies a relatively compact group of mountain blocks and nunataks. In the southern part of the group, flat-topped tabular masses as much as 30 square kilometres in area, bounded by sheer rock walls, rise more than 1000 metres above the general snow level to altitudes of more than 2700 metres. Towards the north the blocky massifs are progressively more dissected, and degenerate into thin ridges and sharp horns, and eventually into disconnected lines of nunataks. Some of these lines of nunataks are crescentic or horseshoe-shaped in plan, enclosing basins as much as 15 kilometres across; and seismic sounding of one basin has shown it to have the form of a great cirque, more than 600 metres deep.

The individual groups of mountain blocks and nunataks within this area have remarkably linear arrangements, each oriented northeast parallel to the main trend of outcrops. The pattern is particularly pronounced at the southern end of the area, where blocky massifs and horn-like nunataks occur in parallel rows, separated by ice-filled corridors that appear to be partly of tectonic origin.

The area of blocky mountains and horns is bounded on the southeast by a trough, nowhere less than 40 kilometres wide, occupied by a great northeast-flowing ice stream whose relatively flat and terraced, but heavily crevassed, upper surface is 200 to 400 metres below the ice level on each side. The upper (southwest) section of the trough has been named Pencksøkka. A seismic profile across it has shown the rock floor to be below sea level in places, as far south as Lat. 72° 50′ S., and about 200 km. inland from the inner edge of the floating ice shelf. The floor of the trough is not smooth, but contains, in profile, several peak-like features that are probably parts of narrow ridges aligned parallel to the long axis of the trough; several of these ridges can be related to belts of crevasses on the surface, and three outcrop as rock islands in the ice stream. There is some evidence to suggest the presence of a considerable thickness of poorly-consolidated material on parts of the trough floor.

The third physiographic unit lies east and southeast of Pencksøkka. The northern part of this unit consists of a nearly continuous series of mountain ranges — Paulsenfjella, Barkleyfjella, Krügerfjellet, Hermannfjella, Gockelkammen — and a group of outlying massifs, Gburektoppane. The ranges rise abruptly as much as 1500 metres above the flanks of the Pencksøkka ice stream and the coastal foreland to the north, and about 500 metres above the rolling surface of the 'inland ice' to the south. The highest known peak is approximately 2830 metres above sea level. In distinct contrast to the blocky form of the mountains northwest of Pencksøkka, these ranges are irregular in plan and profile, characterized by sharp, smooth, sweeping ridges and long curving buttresses, and by a mixture of gentle slopes and steep cliffs that face in almost every direction. The overall pattern of these ranges continues the general northeast trend of the country; but within each mountain group there is no evidence of directional control of the land forms. Across a large north-flowing glacier east of Paulsenfjella and Barkleyfjella, mountains of similar character continue eastward beyond the area investigated.

Southwest from Gockelkammen, this physiographic unit is continued by Neumayerveggen and the southeast wall of Pencksøkka. Neumayerveggen is an irregular, nearly continuous cliff about 50 kilometres long and up to 300 metres high. Its line is continued to the southwest for at least 150 kilometres by an ice-covered escarpment through which groups of nunataks or small mountains project at intervals of 10 to 25 kilometres. From the top of the escarpment, the ice surface rises gently until, 100 kilometres south from the brink, it reaches an altitude of 2700 metres above sea level. A seismic profile shows the escarpment to be not the edge of a great plateau, but rather the crest of a nearly-buried mountain range, similar to that to the northeast, which has dammed back the north-flowing ice. Inland from this range, the underlying rock surface is rolling, and, as far as present knowledge goes, ever more deeply buried by ice.

Almost every mountain massif, and many of the large nunataks, show a rolling upper surface that is quite out of harmony with the steep frost-shattered cliffs and thin, serrated ridges, and is clearly the result of an earlier erosion cycle. Where it is best preserved, this old erosion surface suggests a land form and slope pattern different than that produced by glacier or icecap action; and it seems probable that the land surface, prior to its present occupation by ice, was developed mainly by normal stream action. There is evidence, in the form of erratics and striae, that even the highest mountains have been covered by the general ice sheet; the movement of this ice cover must have been largely controlled by the earlier drainage pattern, and the ice itself must have, in most places, considerably modified the pre-existing land surface. The alpine topography now characteristic of this area must have developed by frost action mainly during and since retreat of the ice to about its present position; and the modern land forms are more or less in harmony with the present-day ice levels.

General Geology.

General Statement.

Two distinct rock assemblages are exposed in western Dronning Maud Land, separated by Pencksøkka and the ice stream flowing north from it to the ice tongue at about the Greenwich meridian.

The nunataks east and south of this ice stream, including Gburektoppane, the range from Paulsenfjella to Gockelkammen, Neumayerveggen and the line of escarpments reaching at least 120 kilometres southwest from it, are composed of a related complex of metamorphic rocks, mainly banded gneisses, amphibolites, schists, and pegmatites. The structure of this complex is variable; over large areas, the foliation of the gneisses is gently-dipping and nearly plane, and in other places deformation has been intense.

A group of small nunataks of granitic gneiss and pegmatite, possibly related to this complex, outcrops among the sediments and sills west of Pencksøkka.

West and north of Pencksøkka and its north-flowing ice stream, the mountains are dominantly a nearly flat-lying series of strongly indurated siltstones, greywackes, and conglomerates, invaded by sills and occasional dykes of dioritic and gabbroic composition. An area in the northeastern part of this region consists of propylitized andesitic flows.

A small mass of similar sediments has been preserved at the southwest end of the southwest wall of Pencksøkka, overlying rocks of the metamorphic complex.

Small basaltic dykes are scattered sparsely throughout the whole area.

Metamorphic Complex.

The metamorphic rocks exposed east and south of Pencksøkka range from chloritic slate and schist to migmatite and granitic pegmatite, and from amphibolite to the most acid gneiss, but about 80 % of the total consists of banded gneisses of acidic to intermediate composition and very simple mineralogy. The most common gneisses are medium-grained, cleanly banded rocks consisting almost entirely of quartz, potash feldspar (mainly microcline), sodic plagioclase (commonly albite-oligoclase $Ab_{94}An_6$ to $Ab_{80}An_{20}$), and biotite. Garnetiferous (grossular) and amphibole-rich gneisses are also widespread; the latter grade into amphibolites, which are ubiquitous as thin layers and pods throughout all the gneisses, and which are the main constituent of a few assemblages as much as 300 metres thick. The common amphibole is a black, relatively iron-rich hornblende, pleochroic green in thinsection; but some rocks contain tremolite-actinolite. Thin layers of biotite-schist, amphibole-biotite-schist, and diopside-amphibolite are fairly common. Minor lithological varieties contain sericite and muscovite, epidote or clinozoisite, or chlorite as essential minerals. Accessory minerals include apatite, sphene, allanite, zircon, magnetite, specularite, and a few metallic sulphides.

The distribution of garnets is unusual. Within some large areas, apparently haphazardly distributed and totalling nearly one-sixth of the outcrop area, rocks of almost all lithological types from acid gneisses to amphibolites are garnetiferous; outside these areas, although the bulk composition, texture, and general mineralogical development appears similar, garnets are rare. Most of the garnets appear as small porphyroblasts, less than 5 mm. diameter, euhedral in the acid rocks, where they are usually surrounded by a zone free of other ferromagnesian minerals, and anhedral in the basic rocks; but well-formed dodecahedrons up to 10 cm. diameter have been found. All garnets tested to date are of the variety grossular, with a refractive index of about 1.75; but the exact composition, and the range of composition in different rocks, has not been determined.

Intercalated with the banded gneisses are many varieties of porphyroblastic gneisses and gneisses with clots and segregated pods and layers of distinctive minerals. Potash feldspar, plagioclase, garnet, amphibole (hornblende), magnetite, biotite, muscovite, and quartz have each been observed as porphyroblasts in different gneisses, but the tendency toward porphyroblastic texture reaches its greatest development in a distinctive and widely distributed, though relatively uncommon, augen-gneiss composed largely of prisms of grey plagioclase (oligoclase) as much as 15 centimetres long. Another unusual and widespread rock type is a fine- to medium-grained quartz-microcline-albite-hornblende-biotite gneiss with conspicuous octahedral porphyroblasts of magnetite 2 to 3 mm. diameter. The rocks characterized by clots and layers of segregated minerals appear to form a complete series from normal banded gneisses with small, irregularly scattered clumps of quartz and feldspar grains, through glomeroporphyroblastic gneisses and gneisses in which most of the acidic material is in irregular, partly-connected pods and lenses, to 'migmatites' in which the original banded rock has been transformed to a poorly foliated, granitic material. In places there appears to be a further development from migmatite to irregular anatectic bodies and pegmatites.

Pegmatitic, and to a lesser extent aplitic, material is associated with the gneiss on nearly every exposure. On many nunataks pegmatites make 2 to 10 per cent of the rock volume. All of the pegmatites are of simple mineralogical composition, comprising the same minerals as the average country rock. Almost all are very acidic in composition, and are composed mainly of quartz and potash feldspar, with greater or lesser amounts of plagioclase and muscovite; biotite and amphibole are, for the most part, minor accessories. Garnets are common in some bodies. Three unusual dykes in Neumayerveggen are composed of grossular, diopside, calcite, and quartz. The texture of most of the bodies is typically pegmatitic, with an intimate intermixture of very coarse (in places more than 1 metre) and fine grains, of euhedral and anhedral outlines. 'Graphic' and myrmekitic quartz-feldspar intergrowths are common. Many of the pegmatites are symmetrically banded, with successive layers differing in texture or composition or both, parallel to their borders. The pegmatites occur mainly as cross-cutting, sharp-bordered bodies of small dimensions: masses larger than 10 metres across are rare. Some of the pegmatites have been emplaced by replacement of the gneisses and amphibolites to form irregular, contorted bodies, or crudely tabular masses and networks that may represent enlarged joints or fracture zones; others are obviously simply fillings of pre-existing fractures. The 'replacement' pegmatites are older than the 'fracture-filling' pegmatites, and usually differ from them by a predominance of white albite-oligoclase over salmoncoloured microcline. In many places there appears to be a complete gradation between the cross-cutting 'replacement' pegmatites and the leucocratic layers in typical banded gneiss, and it is possible that these bodies have developed contemporaneously with the main metamorphism of the rocks, as anatectic differentiates of the surrounding gneisses. Although many bodies are indeterminate, no field evidence of a gradation or direct relationship between the 'fracture-filling' pegmatites and any other rock type has been found.

Aplitic material is relatively rare, and is found mainly in small dykes and stringers associated with, but commonly later than, the 'fracture-filling' pegmatites. In many places a pegmatite-healed fracture has been reopened to admit a thin aplite dyke that splits or borders the pegmatite body for tens of metres.

Rocks of lower metamorphic grade than the common banded gneiss have been found at three places within the area of metamorphic complex mapped. One of these occurences is an isolated nunatak in the Pencksøkka ice stream, entirely composed of low-grade rocks. At the other two places — southwest of Neumayerveggen and in the northern part of Gburektoppane — the low-grade rocks may be traced directly into higher-grade gneisses. The rock types at all three exposures are similar, and include green and grey slates, sericite and sericite-chlorite phyllites, flaky chlorite schists, and relatively pure to micaceous and feldspatic quartzites. Intermediate rock types, probably greater in observed outcrop area than the lowest-grade rocks, include biotite-rich schists and quartzites, light-coloured feldspathic quartzites, and several varieties of fine- to medium-grained rocks that are difficult to classify megascopically, and whose constituents include much epidote and green amphibole. Pegmatite bodies were not observed in these rocks, but quartz veins, commonly containing considerable epidote, chlorite, and pyrite, are ubiquitous.

The structure of the metamorphic complex is very varied. The foliation and banding, on a large scale, outlines fold-like structures that suggest moderate deformation of large thicknesses of rock. Large units may be found oriented at almost any angle, but dips of from 10° to 40° to the southeast or northwest, undulating but fairly constant over areas of 1 to 20 square kilometres, are most common. Outside this range, steep to vertical layering is more common than horizontal layering.

As is common in metamorphic complexes of this type, the smaller-scale structures, as outlined by individual lithological bands and by mineral foliation, are amazingly complex and intricate, with long zig-zag folds, twists, whorls, and crenulations often independent of the structures in nearby bands or layers. This small-scale deformation, at least, has clearly taken place wholly or partly contemporaneously with the metamorphic recrystallization of the main rock layers and the development of the 'replacement' pegmatites, many of which have 'ptygmatic' patterns.

There does not seem to be any evidence of significant overall differential rock movement accompanying or following the emplacement of the 'fracture-filling' pegmatites, which form plane or jagged angular sheets cutting indiscriminately through the intricate contortions of the older rocks. In places there is more than one generation of 'fracture-filling' pegmatites.

A crudely equidimensional body of coarse-grained, mainly non-foliated hornblende syenite, covering about 15 square kilometres, outcrops among the gneisses and amphibolites of Gburektoppane. A similar body has been found on Neumayerveggen, where it has sharp, cross-cutting, intrusive contacts. The syenite varies considerably in composition and texture, but consists mostly of potash feldspar (orthoclase and microperthite) with lesser oligoclase and hornblende.

About 50 kilometres southwest of the west end of Neumayerveggen, numerous bodies of highly sheared or slaty amphibolite-like rock, 50 to 200 metres thick and as much as 3 kilometres long, cut the banded gneisses and amphibolites at various angles. The shearing or foliation of the interior of most of these bodies is parallel to their length, and independent of the attitude of foliation of the surrounding rock. The bodies may represent the remains of dykes that were emplaced prior to the main metamorphism of the complex.

The scanty evidence at present available suggests that these metamorphic rocks belong to a 'shield-type' basement complex which probably underlies much of this part of the Antarctic continent.

Granite Gneiss and Pegmatite Northwest of Pencksøkka.

Four small nunataks at about 72° 46′ S., 06° 45′ W. (approximately 170 kilometres west-northwest of Neumayerveggen) are entirely composed of coarse-grained granitic gneiss and pegmatite. The outcrops, which enclose an area of approximately 100 square kilometres, are very similar in appearance and composition. The dominant rock in each is a light golden-brown-weathering, white to pale buff coarse-grained non-banded gneiss, composed of colourless to smoky-white quartz, light grey to pale buff oligoclase, salmon-coloured microcline, muscovite, and biotite. A small amount of hornblende is present in some places. The gneissic foliation, poor and locally absent, is due mainly to the parallel orientation of mica plates.

Pegmatitic material constitutes about one quarter of the exposed rock in this area, and is found mainly as irregular masses with more or less gradational borders, rather than as discrete dyke-like bodies. The same mineral assemblage is present in the pegmatites as in the surrounding rocks, although several of the coarsest pegmatites are composed almost entirely of one or both of the feldspars, in crystals up to ½ metre long, and quartz.

The whole assemblage closely resembles some of the more granitic portions of the metamorphic complex to the southeast.

Sedimentary Rocks.

The sedimentary rocks exposed north and west of the main area of the metamorphic complex constitute an entirely clastic, typically continental assemblage, varied in detail but essentially uniform over considerable lateral and stratigraphic distances. The commonest rock types form well-bedded repetitive sequences of medium- to fine-grained greywacke, siltstone, and impure sandstone, with minor shale, mudstone, and conglomerate. All of the rocks have been slightly metamorphosed, but most of the original sedimentary textures and structures have been preserved.

About half of the total stratigraphic thickness examined consists of thin beds of greywacke-siltstone. Most of this rock is dull grey, greenish, yellow-grey, or pale rusty brown on weathered surfaces, and grey, green, or yellowgreen where freshly fractured. Most beds are less than 10 centimetres thick. Cross-lamination, mainly on a fairly small scale within individual beds, is very common. Nearly all of these rocks appear to be relatively poor in quartz, and the coarser varieties (particles rarely more than 2 millimetres long) consist largely of broken grains of feldspar and sub-angular fragments of fine-grained, silty rocks in an indeterminate clayey matrix, the whole invariably containing chlorite and one of the epidote-group minerals. For the most part, size sorting within individual beds is excellent, and there is much contrast in grain size between adjacent layers. Some of the thicker (10 to 50 centimetres) beds grade, without distinctive interior lamination, from light-coloured sandy material at the bottom to dark brown shale at the top.

The shales and mudstones interbedded with the greywackes are dark grey, very dark brown to purplish-black, and commonly conspicuously red-weathering. Most are highly indurated, and a great variety of minor structures, such as ripple marks, mud cracks, raindrop impressions, gas pits, etc., are well preserved. The shales occur most commonly as thin partings between greywacke beds, or as irregular groups of beds mainly less than 5 metres thick scattered at rather wide intervals in the greywacke sequence. In a few places shale and mudstone are the dominant rock types in assemblages up to 50 metres thick.

Coarser clastic rocks are common, but make up a small proportion of the total stratigraphic thickness. The most abundant rock in this category is a distinctive and very widespread mudstone-breccia conglomerate consisting of angular fragments of dark grey mudstone in a greywacke and siltstone matrix. True pebble- and boulder-conglomerates are best developed in what appear to be the stratigraphically highest beds of the northeastern part of the area of outcrop of the sediments, where a section nearly 200 metres thick is composed mainly of conglomerate. Elsewhere the conglomerate usually occurs in single isolated beds up to 5 metres thick within thin-bedded greywacke-siltstone assemblages. These conglomerates, distinct from the ubiquitous mudstone-breccia conglomerates, consist of well-rounded pebbles and boulders of greenstone, chloritized volcanic rock or greywacke, white vein quartz, a great variety of cherty and jasperoid material, altered coarsely crystalline rock of dioritic or gabbroic appearance, feldspar porphyry, banded greywacke, siltstone, shale or hornfels, and indeterminate strongly hematitic matter. No pebbles of rock types representative of the metamorphic complex now exposed to the east have been found. The matrix of most of these conglomerates is a grey-green sandy or gritty greywacke. Almost all of the conglomerate beds are very strongly cemented, and joints run through, rather than around, the pebbles. Except in a few beds, only rare boulders are more than 30 centimetres diameter. Sorting is very poor to fair, and there is seldom a conspicuous lamination within the conglomerate layer itself. The beds are less lenticular than many conglomerates elsewhere; thin beds can be followed for many hundred metres without showing appreciable change, and some have been tentatively correlated across as much as 15 kilometres.

In the mountains immediately northwest of Pencksøkka an assemblage about 250 metres thick consists of shales, feldspathic sandstones, and minor pebble-conglomerate heavily impregnated with hematite, forming a conspicuous band of 'red beds' that may be followed for 60 kilometres.

All of the sedimentary beds have suffered slight to moderate metamorphism. The results of this process are manifest in the obviously crystalline or recrystallized character of the coarser greywackes and conglomerate, the hornfelsic appearance, hardness, and splintery fracture of the shales and mudstones, and in the widespread development of epidote, chlorite, and related minerals. The metamorphism has not obliterated the coarser sedimentary characteristics, and features such as lamination, cross-bedding, and ripple-marks have in many cases been emphasized by the recrystallization. The finer textures and structures, including recognizable organic remains, however, have been nearly completely destroyed.

The low-grade regional metamorphism has reached about the same degree of intensity wherever the sedimentary beds have been examined. In addition, sills and larger dykes intruding the sediments have metamorphosed the adjacent beds. This superimposed contact metamorphism is distinctly local and variable, and seldom affects beds more than 10 metres from the contact. Its most conspicuous feature is an irregular bleaching of the contact rocks and a development within them of pods and layers of epidote-group minerals, green amphibole, and feldspar, with lesser amounts of quartz, calcite, specularite, and a variety of minerals whose identity has not yet been determined. The metamorphism has been selective in its development: medium-grained greywacke beds have been most altered, and shales the least affected. Some layers in a finely laminated or cross-bedded series have been almost completely converted to epidote-amphibolite material, while adjacent layers are nearly unaffected.

The sedimentary rocks are almost undeformed; in most parts of the area examined the beds are nearly plane, dipping generally southeast at from 2 to 6 degrees. In a few places whole nunataks have been bodily tilted as much as 30 degrees in different directions.

Toward the southwest end of the line of nunataks forming the southeast wall of Penckshøkka, at 73° 53′ S., 4° 55′ W., a series about 30 metres thick of greywackes and siltstones that appear to be similar to the common sediments to the north lies directly upon the gently-rolling erosion surface of the metamorphic complex.

It is difficult to determine the total thickness of strata exposed. The scarcity of marker beds, the similarity of lithological types throughout the entire assemblage, the prevalence of thick, commonly lenticular sills, and the possibility of block faulting frustrates most attempts to equate beds in scattered groups of nunataks. A careful estimate must await the working out of individual sections and structures. The impression gained in the field is that certainly more than 700, and probably not more than 2000 metres of sedimentary rocks are exposed.

Very little information is available regarding the age of these rocks. The assemblage is essentially conformable in gross structure, but, like any typical continental or shallowwater series, contains innumerable erosional unconformities, any one of which could represent a time lapse of considerable magnitude. On the other hand the lithological similarity of all parts of the section, the frequent repetition of relatively coarse beds, and the fact that the most abundant rock is a greywacke that originally consisted of angular fragments of easily decomposed material, suggests that the entire thickness was deposited at a relatively rapid rate, during a more or less continuous period of erosion of comparatively near-by land masses. (Some of the rocks in the northeastern portion of the area appear to present contradictory evidence: discussion must be deferred until maps and more complete laboratory results are available).

No identifiable fossils could be found in these rocks. A deposit of mud-cracked shale, cross-bedded greywacke, and conglomerate indicates an environment that must have been distinctly unfavourable toward organisms; and any organic remains trapped in the rock have been subject to metamorphism. In several places the mudstone-breccia conglomerate beds look as though they may once have been partly shell beds, but the fragments are invariably too metamorphosed to be recognized as fossils.

Volcanic Rocks.

Small nunataks within an area of about 1100 square kilometres on the west side of the northern part of the ice stream flowing north from Pencksøkka are composed of altered flows of intermediate to basic composition. To the south, these rocks overlie and are in part intercalated with conglomerates and impure quartzites that appear to be part of the main sedimentary assemblage.

The flows are mainly green to dark grey-green, aphanitic to fine-grained rocks. A small proportion are porphyritic, with phenocrysts of ferromagnesian minerals (altered pyroxene) or, less commonly, feldspar. On the basis of their general appearance, it is suggested that most of these flows were originally andesitic. All have been more or less altered, with the development of much chlorite, epidote-group minerals, and amphibolite, to typical greenstones. Most of the flows are amygdaloidal; in places amygdules and gas-cavity fillings make up more than half the rock volume. Amygdules of white and pale blue chalcedonic quartz are most common, but fillings of chloritic or amphibole-like material, and calcite, are locally prominent.

Most of the flows or flow-units appear to be between 3 and 20 metres thick. Flow contacts are commonly conspicuous owing to a thin band of breccia between the flows, and a concentration of amygdules near borders. The upper parts of many flows are characterized by an arrangement of amygdules into strings and bands roughly parallel to the nearby contact, while the lower parts contain a zone of well-developed pipe amygdules about perpendicular to the contact.

The interior of most of the flows is devoid of amygdules.

One nunatak near the northwest border of the area containing exposures of these rocks contains a mass at least 80 metres thick composed entirely of lava pillows.

The nunataks exposing these rocks are too widely scattered to enable a determination of the overall structure to be made. On individual outcrops the flows dip eastward at angles ranging from 5 to 30 degrees.

Dioritic and Gabbroic Sills and Dykes.

Throughout its entire exposed area, the sedimentary assemblage has been invaded by sills. Because of its superior erosional resistance as compared with the sediments, sill material forms the dominant rock of many nunataks. Most of the sills are large; individual thicknesses range up to 200 metres or more, and some single bodies cover an area at least 20 by 10 kilometres.

Individual sills differ in composition and texture, but all appear closely related. The average composition, based solely on megascopic evidence, probably ranges from feldspathic diorite to fairly basic gabbro. Diabasic texture is common. The thicker sills are almost invariably differentiated, with a thin basic or ultrabasic lower zone, a thick central zone of diorite or gabbro, usually medium-grained, and an upper zone, sometimes amounting to one-third of the total thickness, progressively more feldspathic toward a topmost band of highly feldspathic pegmatitic or micropegmatitic material. Most of the sills have chilled upper and lower contacts.

Many interesting and instructive features of differentiation, rhythmic banding, and internal structure are exhibited by the sills. Discussion of these features must be deferred until laboratory investigations have been made.

Most of the sill rocks are relatively fresh, except for a slight saussuritization of the feldspars. In a few places the rocks are epidotized. Fracture surfaces of parts of some sills and the adjacent sedimentary rocks are coated with aggregates of epidote, chlorite or amphibole, quartz, and feldspar, with minor specularite and copper carbonates.

Almost all of the sills are, on a large scale, concordant with the sedimentary beds across any one exposure; but the contacts are invariably transgressive in detail. The transgression of the sill from one group of sedimentary beds to another commonly takes place in a series of steps, with the contact alternately following the bedding planes of thicker, coarser or more siliceous beds, and then breaking directly through finer-grained, less competent beds to the next favourable horizon. In a few places transgressive portions of the sills have violently crumpled the ends of the truncated beds. Such deformation is very local, and has not been observed to extend more than 10 metres from the contact. Inclusions of sedimentary rock within the sills are rare. Those recognized are angular, sharp-bordered, and no more altered than the average contact beds.

Some of the sills are strongly lenticular, or have steep wedge-shaped terminations, over which the sediments are arched.

A few large dykes, up to 50 metres across, similar in composition to the sills, have been found cutting the sediments and sills. They are probably feeders to stratigraphically higher sills. Some of the thicker dykes are composite, or are differentiated between their borders and cores.

Basaltic Dykes.

Small dykes of dark brown-weathering, dark grey to black basaltic rock have been found in all main areas of rock exposures yet visited in Dronning Maud Land. Most are less than one metre thick. They are steep-dipping, commonly angular or irregular in plan, and nowhere numerous — usually several hundred metres or several kilometres apart. With the single exception of five bodies of andesitic aspect in Barkleyfjella, they are of uniform appearance over the whole area. Most of the dykes have chilled (in places nearly glassy) borders and a uniform, aphanitic or very fine-grained interior.

All of the dykes yet examined under the microscope are olivine basalts. Several are porphyritic, with olivine and less commonly pyroxene phenocrysts about one millimetre long, although this feature is not conspicuous in the hand specimens. Vesicular and amygdaloidal varieties have been found, but are rare; the amygdules are sodic plagioclase clinozoisite, apatite, calcite, and sericite.

Tectonics.

Little can be said, at this preliminary stage, regarding the crustal movements to which this portion of the Antarctic continent has been subjected. The metamorphic complex has been involved in what appears to have been a relatively mild orogeny prior to and contemporaneous with its metamorphism. A poorly exposed angular unconformity in Barkleyfjella may be evidence either of two periods of deformation separated by an interval of erosion, or of large scale thrust faulting from the south during a single orogenic cycle. By analogy with the geological history of the Ross Sea section of Antarctica (David & Priestley 1914; Fairbridge 1949) it might be presumed tentatively that the metamorphism of the complex was essentially complete by Cambrian times. Since then the area has been relatively stable. A prolonged period of emergence and erosion was terminated by an epeirogenic sinking of the western portion, at least, of the area into one or more shallow basins or gentle foreland slopes that received a variety of shallow-water and strandflat sediments. There was local volcanic activity during or following the period of sediment accumulation. Following burial (by material of which no trace remains) and slight regional metamophism, the sedimentary assemblage was invaded by numerous large dioritic and gabbroic sills, which have locally domed and differentially lifted the overlying beds. A system of semiparallel, northeast-trending major fractures developed about this time, and may have localized the emplacement of the sills. The last recorded crustal movement has been a general rise of the land, and different blocks, bounded by the northeast-trending fractures, have received slightly different amounts of uplift to form the units from which the present-day mountain ranges have been carved.

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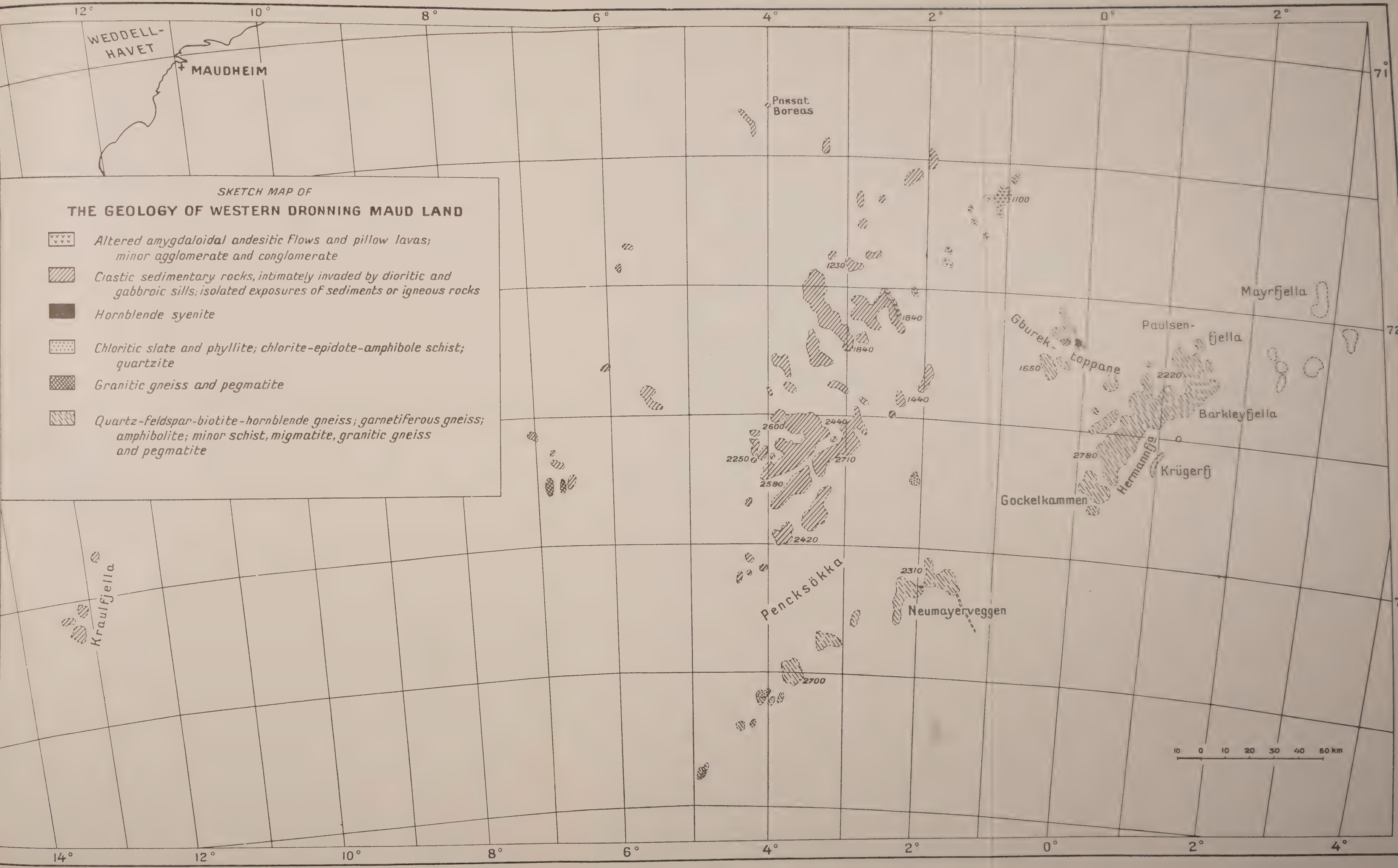

SKETCH MAP OF

THE GEOLOGY OF WESTERN DRONNING MAUD LAND

Altered amygdaloidal andesitic flows and pillow lavas; minor agglomerate and conglomerate
Clastic sedimentary rocks, intimately invaded by dioritic and gabbroic sills; isolated exposures of sediments or igneous rocks
Hornblende syenite
Chloritic slate and phyllite; chlorite-epidote-amphibole schist; quartzite
Granitic gneiss and pegmatite
Quartz-feldspar-biotite-hornblende gneiss; garnetiferous gneiss; amphibolite; minor schist, migmatite, granitic gneiss and pegmatite
WEDDELL-HAVET
MAUDHEIM
Passat
Boreas
1100
1230
1840
1840
1440
1650
Gburek-toppane
Paulsen-fjella
Mayrfjella
2220
Barkleyfjella
2780
Hermannfja
Krügerfj
Gockelkammen
2600
2440
2250
2710
2580
2420
2310
Neumayerveggen
Pencksökka
Kraulfjella
2700
12°
10°
8°
6°
4°
2°
0°
2°
14°
12°
10°
8°
6°
4°
2°
0°
2°
4°
71°
72°
73°
10 0 10 20 30 40 50 km

THE MIDDLE ORDOVICIAN OF THE OSLO REGION, NORWAY

4. Ostracoda.

BY

Gunnar Henningsmoen

(Paleontological Museum, University of Oslo, Oslo 45, Norway.)

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Abstract. 18 paleocop ostracods are described from the Middle Ordovician Ogygiocaris and Chasmops Series of the Oslo Region in Norway. The genus *Piretopsis* is new, and also the species *Ullerella triplicata, U. ventroplicata, Tallinnella trident, T. tumida, T. mjoesensis, T. kiaeri, Ceratopsis? stoermeri, Piretopsis donsi*, and *Eoleperditia skjesethi*. Forms of *Tallinnella* with swollen features are regarded as examples of neoteny as compared with forms with accentuated relief.

Preface.

The present paper is one of several being published by different authors on the Middle Ordovician of the Oslo Region. This team work was initiated by Prof. L. Størmer (Paleontological Institute, Oslo). The writer expresses his gratitude to him for the inspiring way this work has been organized, and for friendly and helpful suggestions during the work. Furthermore the writer wishes to thank him and the other fellow workers, especially cand. real. N. Spjeldnæs and cand. real. S. Skjeseth, for collecting ostracods in the different districts covered by them. It is with pleasure that the writer thanks Miss B. Mauritz for the often troublesome task of photographing the ostracods.

Description of species.

Suborder PALEOCOPA HENNINGSMOEN, 1953

Superfamily Beyrichiacea ULRICH & BASSLER, 1923

Family Tetradellidae SWARTZ, 1936

Subfamily Tetradellinae SWARTZ, 1936

Genus *Tallinnella* ÖPIK, 1937

Type species: — *Tallinnella dimorpha* ÖPIK, 1937.

The genus *Tallinnella* is employed here in the wide sense proposed by HENNINGSMOEN (1953, p. 214), who restricts *Tetradella* to the *Tetradella quadrilirata* group (developing submarginal loculi) and transfers the *subquadrans* group to *Tallinnella*.

Tallinnella trident sp.n.

(Pl. 1, fig. 1, Pl. 2 figs. 1—2.)

Name: — The name *trident* is alluding to the 3-branched pattern of the lobal ridges.

Type data: — Holotype (P.M.O. 66420) is an external impression of a left valve from 4a α_2, collected at Muggerudkleiva, Sandsvær by L. STØRMER, 1925.

Material: — A great number of external and internal impressions of separated valves preserved in shale.

Diagnosis: — *Tallinnella* species with L1, L3, and L4 rather narrow and forming a 3-branched pattern. L4 does not reach the dorsal border. L2 is small and node-like.

Description: — Adult valves are up to 2,5 mm long and 1.5 mm high. Outline with slight forward swing. Surface with high relief. L1, L2, and L3 are narrow, almost ridge-like, and form a 3-forked pattern. L1 and L3 project slightly above the dorsal margin, whereas L4 fades away well below the dorsal margin. L2 is small and node-like and gives the impression of being situated on the inner slope of L1. S2 and S3 deep and wide, S1 faint and narrow. The lobes are separated from the velate frill by a well defined groove. Velate frill not very wide but convex; raising rather steeply up from the carapace wall and bending rather abruptly down again; most convex along the anterior margin. The frill is restricted, and does not reach more than half way up along the posterior margin. The material is not too well preserved and no dimorphism has been detected in the frill.

Relationships: — *Tallinnella trident* sp.n. may be a close relative of species like *T. marchica* (KRAUSE, 1889) and *T. dimorpha* (ÖPIK, 1937) but differs from these in the almost ridge-like appearance of L1, L3, and L4 and the weak development of L2. In this respect *T. trident* sp.n. resembles *T. scripta* (HARPER, 1947), which differs, however, i.a. in the long L4 (reaching the dorsal border) and in having a ridge between the ventral end of L3 and the velate ridge. *T. tumida* sp.n. may be closely related to *T. trident* sp.n.

Occurrence: — Associations of numerous specimens in 4a α_1—4a α_3. Eiker-Sandsvær (Muggerudkleiva, Ravalsjø-elva, etc.), Modum (Melåen).

Tallinnella tumida sp.n.

(Pl. 1, fig. 2, Pl. 2 figs. 3—4).

Name: — The name *tumida* is given due to the somewhat swollen appearance of this species.

Type data: — Holotype (P.M.O. 66421) is a left valve from 4a α_3, collected from the section Bjerkås—Djuptrekkodden, Røyken, by N. Spjeldnæs, 1951.

Diagnosis: — A *Tallinnella* species with rather swollen lobes, confluent with the extralobal area. L2 small and node-like, giving the impression of being situated on the posterior slope of L1. Frill narrow and restricted.

Description: — Adult valves are up to about 2.5 mm long and 1.5 mm high. Outline with slight forward swing. Surface rather convex, with swollen relief. The lobal area is confluent with the extralobal area. L1 and L3 project slightly above the hinge line and bound a semi-elliptical depressed area, consisting of S1, L2, and S2. L2 is small and node-like and situated, as it appears, on the posterior slope of L1, and is not as high as L1 and L3. L3 rises well above L4, S3 being developed almost as a semisulcus. A narrow groove is formed where the velate frill rises from the valve wall. The frill is narrow, somewhat convex, and restricted to the anterior and ventral margin. Surface (including frill) appears smooth.

Relationships: — *Tallinnella tumida* sp.n. may be compared with *T. marchica lata* (KRAUSE, 1891). The figures of the latter show several differences in details from *T. tumida* sp.n., but it is not impossible that a better knowledge of *T. marchica lata* may prove that these forms are synonyms. *T. tumida* sp.n. agrees with *T. trident* sp.n. in several features as type and position of L2, restricted frill, etc. and differs mainly in having more swollen features. It is probale that *T. tumida* sp.n. has developed from *T. trident* sp.n.

Material: — A few separated valves preserved in limestone.

Occurrence: — In 4a α_3 of the section Bjerkås—Djuptrekkodden, Røyken.

Tallinnella mjoesensis sp.n.

(Pl. 1, fig. 3, Pl. 3 figs. 1—5).

Name: — From Lake Mjøsa, as this ostracod occurs in the Mjøsa districts.

Type data: — Holotype (P.M.O. 66422) is a right valve from the Cyclocrinus Beds at Furuberget, Nes—Hamar district, collected by S. Skjeseth, 1950.

Material: — About 50 valves and impressions of these, including larval instars.

Diagnosis: — A *Tallinnella* species with narrow, ridge-like lobes. The ventral connecting ridge is, however, poorly developed. Velate structure ridge-like, not quite reaching the dorsal margin in either end.

Description: — Adult valves about 2.5 mm long and 1.5 mm high. Ends subequal, anterior end slightly more pointed than posterior end. Surface with very high relief. The lobes are developed as long, high and narrow ridges. They are connected ventrally, but the connecting ridge is poorly developed. L2 and L3 are higher than L1 and L4, L3 being the higher. L1 is curved parallel to the anterior margin and is somewhat overhanging outwards. L2 is almost straight, with the dorsal end only slightly bulbous, and shorter than L1 and L3. L1 is slightly oblique and reaches farther up dorsally than the other ridges; but does not quite reach the dorsal border. L4 is curved parallel to the posterior border. The sulci are rather flat, as is the groove between the lobal area and the velate ridge. The latter is ridge-like, high and narrow as are the lobes. It is set off from the valve wall at a rather high angle. The posterior end is the highest part of the velate ridge. The velate ridge does not reach the dorsal border at either end; the posterior end reaching slightly farther up than the anterior end. Surface of the valve (including velate ridge) appears smooth. Left valve with a fine ridge along the dorsal contact margin, fitting into a corresponding fine groove in the right valve. Free border of the left valve slightly overlaps the right valve.

Larval instars. Small valves (1.5 mm long and less) have a more or less even convex surface, only L2 and L3 being set off as two not very well defined small nodes. In a valve 1.75 mm long the two other ridges are developed as well, and the lobal area separated from the velate ridge, which is swollen. The posterior lobes are ridge-like, whereas the anterior lobe (L1) is still swollen, and hardly separated from the velate ridge. In larger valves the anterior lobe and velate ridge assume the shape of the adult.

Relationships: — The long L2 reminds one of the poorly known *T. digitata separata* ((Steusloff, 1894) and also early *Tallinnella* species as *T. grewingki* (Bock, 1867). *T. mjoesensis* is probably close to *T. kiaeri* sp.n. (cf. p. 41).

Occurrence: — Rather common in the Cyclocrinus Beds and Mjøsa Limestone at Furuberget (Hamar district).

Tallinnella kiaeri sp.n.

(Pl. 1, fig. 4, Pl. 4 figs. 1—6).

1897 *Beyrichia* cf. *Marchica* Krause—Kiær, p. 39 (Listed from "5a", i.e. Mjøsa Limestone).

1909 *Beyrichia* aff. *Marchica* Krause—Holtedahl, p. 32 (Listed from Cyclocrinus Beds).

Name: — This species is named after its finder, Johan Kiær.

Type data: — Holotype (P.M.O. 66432) is a left valve from the Mjøsa Limestone at Furuberget (Hamar district), collected by S. Skjeseth, 1950.

Material: — More than hundred separated valves, including larval instars.

Diagnosis: — A *Tallinnella* species with L1 and L4 more or less completely fused with the swollen velate ridge. L2 and L3 are developed as swollen lobes, hardly connected ventrally and not connected with the velate ridge.

Description: — Adult valves are up to 3.9 mm long and 2.1 mm high. Ends subequal or with slight forward swing. Surface not very convex but with swollen structures. L1 and L4 swollen and more or less completely fused with the swollen velate ridge (see Pl. 4, fig. 1). The submarginal ridge thus formed is subparallel to the free border, except for its posterior end which turns round and points forwards. L2 and L3 form swollen lobes, not connected ventrally with the velate ridge. There is considerable variation in the outline of this species, as in the form of the lobes. L2 and L3 may be short or almost reach the velate ridge and sometimes they are indistinctly united ventrally to form a U-shaped loop.

Larval instars. Larval valves are more tumid and with less well developed lobation. In the smallest valves only L2 and L3 can be distinguished from the general convex surface.

Relationships: — As discussed below, *T. mjoesensis* may be a rather closely related species. *T. kiaeri* sp.n. differs rather much in general appearance from other *Tallinnella* species, recalling such genera as *Drepanella* and *Bollia*. If *T. kiaeri* is related to *Drepanella*, with which it also agrees in attaining a large size, this would mean that *Drepanella* had developed from *Tallinnella*. It is, however, possible that the pattern of *T. kiaeri*, *Drepanella bigeneris* ULRICH, 1894, and *Bollia* may have developed in different stocks.

Occurence: — Rather common in Mjøsa Limestone at Nes (Helgøya) and Toten (Einavatnet).

Swollen forms versus forms with accentuated relief in Tallinnella.

The differences between *Tallinnella trident* and *T. tumida* may be ascribed to the latter being more convex and having swollen lobes. There is a similar difference between *T. marchica marchica* and *T. marchica lata*. It is interesting that forms with accentuated relief have larval instars with swollen features. Swollen adult forms may be regarded as examples of neoteny as compared with forms with accentuated relief. It is possible that this is due to environmental conditions or mode of habitat and that the swollen forms need not be genetically distinct from those with accentuated relief. *T. kiaeri* sp.n. is probably specifically distinct from *T. mjoesensis*, but they are probably closely related. *T. kiaeri* corresponds well to larval instars of *T. mjoesensis* and may be regarded as a neotenous species.

Genus *Ceratopsis* ULRICH, 1894

Type species: — *Beyrichia chambersi* MILLER, 1874

Ceratopsis? sp.

(Pl. 1, fig. 5, Pl. 2, fig. 5).

Material: — A right valve and a left valve with the test missing except for marginal parts. Preserved in limestone.

Description: — The two valves are about 2.5 and 1.5 mm high (including frill). Well developed relief. Outline with almost subequal ends. L1 broken off in both specimens but apparently forming a horn, whose base is round. L2 forms a small isolated node. L4 and especially L3 are well developed and long, reaching the velate frill. The frill is wide and restricted to the anterior and ventral margin and is radially striated. The test is granulated where it is preserved (along the frill).

Occurrence: — Rare in Coelosphaeridium Beds at Ringsaker (Skjønsby).

Ceratopsis? stoermeri sp.n.

(Pl. 1, fig. 6, Pl. 2, fig. 6).

Name: — This species is named in honour of Prof. Leif Størmer, who collected it, and who initiated the detailed study of the Middle Ordovician of the Oslo Region.

Type data: — Holotype (P.M.O. 60402) is an external impression of a right valve from 4a α_2, collected by L. Størmer in 1925 at Muggerudkleiva, Sandsvær.

Material: — External impressions of two left and two right valves, all more or less damaged and preserved in shale.

Diagnosis: — A tetradellid species with *Ceratopsis*-like horn formed by L1 and probably also L2. The ventral connecting ridge of the lobes does not quite reach the horn. Velate frill wide and apparently restricted.

Description: — The four valves are rather small, c. 1.3 mm long and 0.8 mm high (including frill). Outline with forward swing. Well developed relief. L1, possibly with the incorporated L2, forms a comparatively large horn with round base. It points out from the valve wall. L3 and L4 are connected ventrally, and the connecting ridge continues forwards subparallel to the free border but does not quite unite with the horn. Frill rather wide, radially striated and apparently restricted to the anterior and ventral margin.

Relationships: — The horn of *Ceratopsis*? *stoermeri* sp.n. has a slightly more posterior position than is usual in *Ceratopsis* and may include L2. Otherwise the general appearance of the lobes recalls *C. oculifera* (HALL, 1871), where L2 also seems to be more or less incorporated in the horn. Young forms of *C. oculifera* are rather like the Norwegian form. *C. oculifera* and other American species of the genus seem to differ from *C.*? *stoermeri* sp.n. in having a narrow and entire velate frill. *C.*? *stoermeri* sp.n. may possibly be related to *Piretopsis donsi* gen. & sp.n.

Occurrence: — Rare in 4a α_2 at Sandsvær (Muggerudkleiva).

Subfamily Piretellinae ÖPIK, 1937
Piretopsis gen.n.

Name: — The name suggests the likeness to *Piretella* ÖPIK, 1937

Type species: — *Piretopsis donsi* gen. & sp.n.

Diagnosis: — See diagnosis of the type species (only known species).

Relationships: — *Piretopsis* gen.n. shares the U-shaped ridge of L1 and L3 with *Piretella* ÖPIK, 1937 and *Rakverella* ÖPIK, 1937 (included in the Piretellinae by HENNINGSMOEN, 1953). Furthermore the velate frill is restricted and curved as in some individuals (female type) of the Piretellinae. On the other hand *Piretopsis* is undoubtedly close to the Tetradellinae, from which the Piretellinae may have developed (HENNINGSMOEN, 1953). Middle Ordovician members of the Tetradellinae often have a restricted frill and thus agree with the Piretellinae, in contrast to Lower Ordovician Tetradellinae, whose velate structure is entire. *Piretopsis* has a *Ceratopsis*-like horn. This does not necessarily involve it having developed from *Ceratopsis*, as this type of horn may have developed in different stocks (cf. HENNINGSMOEN, 1953, p. 216). *Ceratopsis*? *stoermeri* sp.n. may be related to *Piretopsis*. *Piretella* ÖPIK, 1937 may have developed from *Piretopsis* or similar forms through effacement of lobation and loss of C4. Morphologically, at least, *Piretopsis* is intermediately between the Tetradellinae and *Piretella*.

Piretopsis donsi gen. & sp.n.

(Pl. 1, fig. 7, Pl. 2 figs. 7—8).

Name: — This species is named after its finder, Curator J. A. Dons.

Type data: — Holotype is the external impression (P.M.O. 66424) and internal mould (66425) of a right valve, collected in 4a β at Ullern, Oslo by J. Dons, 1948. It is the only specimen known and is preserved in weathered limestone.

Diagnosis: — A tetralobate species with a U-formed crest formed by C1 and C3. Another crest (C4) is not united with this. L1 is horn-like. Velate frill restricted and curved in the only specimen known (probably dimorphic).

Description: — Length 2.3 mm, hight 1.6 mm. Outline with subequal ends. Moderate relief, except for the horn and crests. Tetralobate. L1 forms dorsally a horn-like projection, pointing outwards and slightly backwards. L2 very small and node-like. L3 well defined and higher than the less well defined L4. L1, L3, and L4 bear crests. C3 runs along the anterior edge of L3 and unites ventrally with C1 to form a U-shaped ridge. C4 runs along the anterior edge of L4 and continues forwards to below the median sulcus, but does not unite with the C1—C3-crest. The lobate area is confluent with the extralobal area. The velate frill is wide and restricted and apparently radially striated. It is rather convex and may represent the female type. Surface (excluding frill) is densely covered with small granules (possibly spine bases). Along the dorsal margin there is a row of coarser granules, probably corresponding to the dorsal crest in *Piretella.*

Occurrence: — The specimen was found in 4a β at Skogfaret, Ullern, Oslo.

Subfamily Bassleratiinae SCHMIDT, 1941.

Genus *Steusloffia* ULRICH & BASSLER, 1908.

Type species: — *Beyrichia costata* LINNARSSON, 1869.

Steusloffia costata (LINNARSSON, 1869).

(Pl. 1, fig. 8).

1869 *Beyrichia costata* n.sp. — LINNARSSON, p. 85, pl. II, figs. 67—68.
1889 *Strepula Linnarssoni* n.sp. — KRAUSE, p. 16, pl. II, figs. 4—5.
1908 *Beyrichia* (*Steusloffia*) *linnarssoni* (KRAUSE) — ULRICH & BASSLER, pp. 282, 296, text figs. 8, 34, pl. 38, fig. 1.

1923 *Steusloffia linnarssoni* (Krause) — Ulrich & Bassler, 1923, p. 308, text fig. 5.
1924 *Steusloffia* (*Strepula*) *Linnarssoni* Kr. — Kummerow, p. 412, text fig. 1.
1934 *Steusloffia linnarssoni* (Krause) — Bassler & Kellett, p. 475, text fig. 9:5.
1937 *Steusloffia linnarssoni* (Krause) — Öpik, p. 50, text fig. 7 (Compared with *S. rigida* Öpik, 1937).
?1937 *Steusloffia multimarginata* n.sp. — Öpik, 1937, p. 52, pl. IV, fig. 4.
1940 *Steusloffia costata* (Linrs.) — Thorslund, p. 176, pl. 2, figs. 10—12; pl. 3, fig. 5; text fig. 58. (Claims that *S. linnarssoni* is a synonym and *S. multimarginata* probably merely a variety. Describes larval instars.)
1948 *Steusloffia costata* (Linrs.) — Thorslund, p. 368, pl. XX, fig. 10.

Type data: — As lectotype is here chosen the specimen figured by Linnarsson (1869, pl. II, fig. 67) from »Beyrichienkalk« in Västergötland, Sweden.

Remarks: — The Norwegian form is similar to the Swedish form as described and figured by Thorslund (1940).

Horizon in Norway: — 4a β—4b α, thus probably not also 4a α as indicated by Dons & Henningsmoen (1949, p. 30).

Localities: Oslo—Asker (Bygdøy, Gaustadskogen, Ullernåsen, Asker), Ringerike (Stubbdal) — Sweden, Germany (Scandinavian drift boulders), ?Esthonia.

Familiae incerta

Genus *Ullerella* Henningsmoen, 1950 (= *Ullia* Henningsmoen, 1949 non Roewer, 1943)

Type species: — *Ullia ulli* Dons, 1949.

Remarks: — The taxonomic position of this genus is uncertain. The triangular cross section of the carapace is shared with many Sigmoopsiidae, and young instars of *Ullerella* recall *Bolbina* Henningsmoen, 1953. *Ullerella* does not, however, show the sigmoidal median sulcus or the carinal dimorphism typical for this family and seems to have a better developed velate structure. In these features it corresponds better to the Tetradellidae, and it may be best for the present to regard the family reference as unsettled. The Silurian zygobolbids remind one of *Ullerella*. They differ i.a. in developing a dimorphic pouch or swelling of the carapace wall. As the Zygobolbinae (and thus the Beyrichiidae, cf. HENNINGSMOEN, 1953, p. 236) no doubt have developed from Ordovician Beyrichiacea, which are not known to develop a similar pouch, it is not impossible that they have developed from *Ullerella* or related forms. In *Ullerella holtedahli* (DONS, 1949) and *U. triplicata* sp.n. the inner U-shaped ridge shows a characteristic widening in the ventral half of the posterior part of the ridge. It is interesting to note that a similar widening is seen in in the U-shaped ridge of some zygobolbids, especially in *Mastigobolbina*. As long as no intermediate forms are known between the Middle Ordovician *Ullerella* and the Silurian zygobolbids, the relationships between these may only be regarded as a suggestion.

Ullerella triplicata sp.n.

(Pl. 1, fig. 9, Pl. 5, figs. 5—6).

Name: — The name alludes to the three sets of ridges in this species.

Type data: — Holotype (P.M.O. 66426) is the external impression of a left valve, collected in 4a α_2 at Muggerudkleiva, Sandsvær by L. Størmer, 1925.

Material: — A large number of external impressions and internal moulds of separated valves, preserved in shale.

Diagnosis: — An *Ullerella* species with 3 concentric U-shaped ridges, reaching the dorsal margin in adult specimens.

Description: — Adult valves c. 1.6 mm long and 1.0 mm high. Outline with forward swing. Sulcus (S2) long and wide, bent slightly forwards ventrally. The sulcus is surrounded by an elevated area consisting of 3 concentric U-shaped ridges divided by two narrow grooves. The inner ridge is rather wide and lobe-like, except ventrally, where it narrows to the same width as the thin outer ridges. All ridges reach the dorsal border. The outermost ridge slightly overhangs the free border ventrally. The surface appears smooth.

Larval instars. In small specimens the posterior halves of the U-shaped ridges do not reach more than half-way up between the ventral and dorsal border.

Occurrence: — In 4a α_1, 4a α_2, and 4a α_3? at Sandsvær (Muggerudkleiva, Ravalsjøelva, etc.) and Modum.

Ullerella ventroplicata sp.n.

(Text fig. 1, Pl. 1, fig. 10, Pl. 5, figs. 3—4).

Name: — The name indicates that the ridges are mainly restricted to the ventral half of the valve.

Type data: — Holotype (P.M.O. 4180 a) is a right valve, collected in 4a α_4 at Gomnes, Ringerike by J. Kiær, 1915.

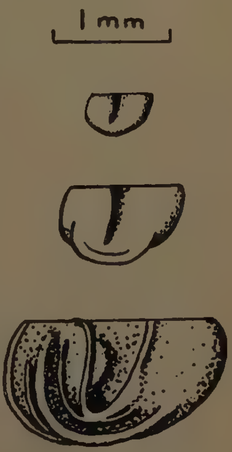

Fig. 1. *Ullerella ventroplicata* sp.n. Two larval and an adult valve.

Material: — Some 10 separated valves (including larval instars) preserved in limestone.

Diagnosis: — An *Ullerella* species with 3 concentric ridges. Only the inner ridge is U-shaped and reaches the dorsal margin (in adult specimens). A fourth ridge is developed anterior to the anterior branch of the outer ridge.

Description: — Adult valves c. 1.9 mm long and 1.2 mm high. Outline with forward swing. The sulcus (S2) is well developed, curving somewhat forwards ventrally. It is surrounded by a U-shaped ridge, reaching the dorsal border in both ends. Two more concentric ridges lie outside this ridge; they do not reach the dorsal border, but reach somewhat further up anteriorly than posteriorly. A fourth ridge is developed parallel to and outside the anterior branch of the third ridge. A faint lobe is developed beneath the anterior branch of the inner ridge. The two inner ridges are coallesced ventrally. Surface is finely granulated, except for the ridges which are smooth.

Early larval instars (text fig. 1). C. 0.5 mm long valves show the general form and lobation of adult valves, but the ridges are hardly traceable. In c. 1 mm long valves two ridges are clearly visible in the ventral half. The innermost ridge of the adults is not developed. *Later instars*. In up to 1.5 long valves the ridges are still confined solely to the ventral half of the valve. The innermost ridge is still present rather as lobes, but fused ventrally with the second ridge (from the sulcus) as in the adults. The anterior branches of the two ridges (in the adult ridge 2 and 3 from the sulcus) extend gradually further upwards into the dorsal half of the valve. The anterior branch of the outer ridge splits into two ridges.

Occurrence: — In 4a α_4 at Oslo—Asker (Tøyen, Gamlebyen, Engervik) and Ringerike (Gomnes).

Ullerella ulli (Dons, 1949).

(Pl. 1, fig. 11).

1949 *Ullia ulli* Dons sp.n. — Dons & Henningsmoen, p. 28, pl. 1, figs. 1—4.
1950 *Ullerella ulli* (Dons, 1949) — Henningsmoen, p. 244.

Type data: — Holotype (P.M.O. 65827) is an external impression of a right valve collected in 4a β at Abbediengkollen, Ullern, Oslo by J. Dons, 1943.

Diagnosis: — An *Ullerella* species with two concentric U-shaped ridges, both reaching the dorsal border in both ends.

Occurrence: — This species was originally stated to occur in beds believed to belong to 4a α—β. A closer study of these beds has revealed that they are restricted to 4a β. Oslo—Asker (Oslo, Ullern, Bygdøy), Ringerike (the section 4a β—4b α at Kullerud, 50 m above the base), Gjerpen—Langesund (Frierfjorden).

Ullerella holtedahli (Dons, 1949).

(Pl. 1, fig. 12).

1949 *Ullia holtedahli* Dons sp.n. — Dons & Henningsmoen, p. 29, pl. 1, figs. 5—8.

Type data: — Holotype (P.M.O. 65828) is an external impression of a left valve from 4b α at Øvre Ullern Terrasse, Ullern, Oslo and collected by J. Dons, 1943.

Diagnosis: — An *Ullerella* species with 2 concentric U-shaped ridges. Their ends reach the posterior border, except the posterior branch of the outer ridge, which fades away after a short distance.

Additional description: — A fine striation is sometimes seen on the ridges in their longitudinal direction.

Occurrence: — In 4b α (possibly also in 4b β); Oslo—Asker (Tøyen, Grefsen, Tørtberg, Bygdøy, Ullern, Asker) and Ringerike (Kullerud, Gomnes—Rud).

Family Sigmoopsiidae Henningsmoen, 1953

Subfamily Sigmoopsiinae Henningsmoen, 1953

Genus *Bolbina* HENNINGSMOEN, 1953

Type species: — *Entomis variolaris* BONNEMA, 1909.

Bolbina variolaris (BONNEMA, 1909)

(Pl. 1, fig. 13).

1909 *Entomis variolaris* n.sp. — BONNEMA, 1909, p. 68, pl. 5, figs. 10—11.
1909 *Entomis oblonga* var. *Kuckersiana* n. var. (pars) — BONNEMA, p. 67, pl. 5, fig. 9.
1934 *Ctenobolbina variolaris* (BONNEMA) — BASSLER & KELLETT, p. 255.
1937 *Ctenobolbina variolaris* (BONNEMA) — ÖPIK, pl. XI, fig. 11.
1940 *Winchellatia variolaris* (BONNEMA) — KAY, p. 254, pl. 32, fig. 12.
1940 *Ctenobolbina variolaris* (BONN.) — THORSLUND, p. 173, pl. 3, figs. 15—17.
1941 *Winchellatia variolaris* (BONNEMA) — SCHMIDT, p. 36.

Type data: — As lectotype is here selected a right valve figured by BONNEMA (1909, pl. V, fig. 11) from the Kukruse formation in Esthonia.

Present material: — A number of separated valves preserved in limestone. The largest specimen reaches a length of 1.4 mm.

Remarks: — As remarked by BONNEMA (1909) and THORSLUND (1950) this species is dimorphic. It is a carinal dimorphism, typical of the Sigmoopsiidae (cf. HENNINGSMOEN, 1953, p. 201).

Occurrence: — In 4b δ of Oslo—Asker (Terneholmen, Sylling). Lower Chasmops limestone in Sweden, and Kukruse (Kúckers) formation in Esthonia.

Family Primitiidae ULRICH & BASSLER, 1923

Genus *Primitia* JONES & HOLL, 1865

Type species: — *Beyrichia strangulata* McCOY, 1852.

Primitia suecica (THORSLUND, 1948)

(Pl. 1, fig. 14, Pl. 5, fig. 7).

1948 *Eurychilina suecica* sp.n — THORSLUND, 1948, p. 368, pl. XX, figs. 8—9.

Type data: — The type material is from a core through the Lower Chasmops Series at Kinnekulle, Vestergötland, Sweden.

Present material: — A few external impressions and internal moulds of separated valves, preserved in shale.

Description of present material: — Valves up to 1.5 mm long and 1.1 mm high. Outline with slight forward swing. Sulcus (S2) short, straight, almost normal to the hinge line and well defined, except dorsally where it widens out. Surface otherwise rather evenly convex. A velate frill is restricted to the anterior and ventral parts of the posterior margin. It is radially striated. A sausage-like velate pouch is developed in some specimens. The pouch appears to run along the ventral and half way up the anterior margin. Free border appears spinose, at least posteriorly. Surface granulated. A row of larger granules (spines?) near the dorsal margin. It may correspond to the dorsal ridge in some *Primitia* species.

Relationships: — This form may be closely related to the type species, *Primitia strangulata*, but the latter appears to have a shorter velate pouch and differs also in other minor details.

Occurrence: — 4b γ Oslo—Asker (Hvalstad). — Lower Chasmops Series in Sweden (Vestergötland).

Family Eurychilinidae ULRICH & BASSLER, 1923
Subfamily Eurychilininae ULRICH & BASSLER, 1923
Genus *Platybolbina* nomen nov.

Remarks: — This name is given here for *Platychilina* THORSLUND, 1940, p. 169 (non KOKEN, 1892, as pointed out by AGNEW, 1944) *Platychilina* KUMMEROW, 1933, p. 45 is a nomen nudum (cf. HESSLAND, 1949, p. 253).

Type species: — *Primitia distans* KRAUSE, 1889 (Pl. 1, fig. 3).

Diagnosis: — (According to THORSLUND, 1940, p. 169); Similar to *Chilobolbina*, but lacking a well-defined sulcus, instead having a conspicuous, subcentral muscle spot.

Platybolbina kapteyni (BONNEMA, 1909)
(Pl. 1, fig. 15).

1896 *Primitia distans* KRAUSE — KRAUSE, p. 933, pl. 25, figs. 7—8.
1909 *Primitia Kapteyni* n.sp. — BONNEMA, p. 29, pl. 6, fig. 31.
1923 *Chilobolbina kapteyni* — ULRICH & BASSLER, p. 516.
1934 *Chilobolbina kapteyni* (BONNEMA) — BASSLER & KELLETT, p. 245.
1940 *Platychilina kapteyni* (BONN.) — THORSLUND, p. 169, pl. 1, figs. 8—10; pl. 5, fig. 7.

Type data: — Holotype is the right valve figured by BONNEMA (1909, pl. 6, fig. 31) from the Kukruse (Kuckers) formation in Esthonia.

Present material: — A few separated valves preserved in limestone and external impressions and internal moulds preserved in shale.

Description of present material: — Length c. 1.6 mm, height c. 1.2 mm. A faint depression takes the place of the median sulcus. In front of this, at least in internal moulds, may be seen a small and faint lobe (L2). The velate frill runs along all of the free border, except for a short distance near the posterior dorsal corner. It is dimorphic; in some individuals the greater part of the frill is convex, forming a velate pouch. The frill is radially striated. The surface ornamentation of the valve gives a reticular appearance, with the exception of a subcentral spot and of the sulcoid depression. Furthermore the cardinal corners may be devoid of ornamentation in a triangular area, as described by KRAUSE (1896, p. 933) from the posterior cardinal angle and by BONNEMA (1909, p. 29).

Occurrence: — 4b γ (and 4b δ ?) at Oslo—Asker (Hvalstad).

? Subfamily Euprimitiinae HESSLAND, 1949

As discussed by HENNINGSMOEN (1953, p. 228) this subfamily is only tentatively assigned to the Eurychilinidae.

Genus *Euprimites* HESSLAND, 1949

Type species: — *Euprimites reticulogranulata* HESSLAND, 1949.

Euprimites ? *suecica* (THORSLUND, 1949).

(Pl. 1, fig., 16, Pl. 5, fig. 2).

1940 *Ctenobolbina suecica* n.sp. — THORSLUND, p. 172, pl. 2, figs. 13—14, pl. 3, fig. 14.
1940 *Euprimitia minor* n.sp. — THORSLUND, p. 164, pl. 1, figs. 11—12.
1949 *Euprimites minor* (THORSLUND, 1940) — HESSLAND, p. 250.

Type data: — The holotype is from Lower Chasmops Limestone in the Lockne area, Sweden, and was figured by THORSLUND (1940, pl. 2, fig. 14).

Present material: — A number of separated valves (including larval instars) preserved in limestone.

Remarks: — The Norwegian material shows the same type of velate dimorphism as the Swedish form; whereas some individuals have only a velate ridge (male type), others have a shelter-like frill along the anterior two thirds of the ventral border. In young specimens the velate ridge is less well developed and in the smallest specimens the velate structure appears to be absent. Small specimens with a slightly developed velate ridge agree quite well with *Euprimitia minor*, which are believed by the writer to be larval instars of *Euprimites? suecica.*

Relationships: — *Euprimitia minor* was transferred to *Euprimites* by HESSLAND (1949, p. 250). The Lower Ordovician type species of *Euprimites* shows no dimorphism as in *E.? suecica*. This may be due to the rather small material of the type species (2 carapaces and 2 valves). If more material will later prove that the type species shows no velate dimorphism, *E.? suecica* and allied species should probably not be included in this genus, and are here only tentatively assigned to it. *E.? suecica* is most probably not related to *Ctenobolbina* (cf. HENNINGSMOEN, 1953, p. 229). *Euprimites? ("Euprimitia") plena* (ÖPIK, 1937) appears to be a closely related species, and also *Euprimites? ("Primitia") eutropis* (ÖPIK, 1937). *Haploprimitia inconstans* ÖPIK, 1937 is probably based on larval instars of *Euprimites? eutropis.* All the above mentioned species tentatively assigned to *Euprimites* may be related to *Primitia*. Further study is however needed, to trace the relationships between these species, the type species of *Euprimites*, and the genera *Euprimitia* and *Primitia*. If they are all closely related, the Euprimitiinae and probably also the Eurychilinae may again be included in the Primitiidae. As discussed by HENNINGSMOEN (1953, p. 226), the *Primitia* and closely allied species may also possibly have developed from the Piretellinae. For this reason the family Primitiidae was restricted to *Primitia* and closely allied, and the Euprimitiinae and Eurychilininae included in a separate family Eurychilinidae.

Occurrence: — 4b δ in Oslo—Asker (Terneholmen, Sylling).

Superfamily Leperditiacea BASSLER & KELLETT, 1934
Family Leperditiidae JONES, 1856
Subfamily Leperditiinae JONES, 1856
Genus *Eoleperditia* SWARTZ, 1949

Type species: — *Cytherina fabulites* CONRAD, 1843.

Eoleperditia skjesethi sp.n.
(Pl. 1, fig. 17, Pl. 5, fig. 1).

Name: — This species is named in honour of cand. real. Steinar Skjeseth, who has thoroughly studied the stratigraphy of the Mjøsa districts and who has collected this, as well as several others of the ostracods described in this paper.

Type data: — Holotype (P.M.O. 66427) and only specimen known; a left valve preserved in limestone from the Mjøsa Limestone, and collected at Bergvika, Helgøya, Nes—Hamar district by S. Skjeseth.

Diagnosis: — A leperditiid about 1.6 times as long as wide, with no chevronmark but with subovoid adductor muscle scar with some 100 angular muscle spots.

Description: — The holotype is 6.5 mm long and 4.0 wide. Shell subovate in outline, truncated dorsally by straight hinge which extends for about four-sevenths of the greatest length. Anterior cardinal angle less obtuse than posterior cardinal angle. Surface of valve moderately convex with the greatest convexity just behind the muscle scar. Anterior corner slightly depressed, as well as posterior margin. The adductor scar is subovoid and with some 100 angular muscle spots. No other muscle spots or venose lines are seen. The eye tubercle is small and lies just in front of the upper end of the muscle scar.

Relationships: — As only an external left valve is known, it is uncertain whether this species shows any of the prongs described in the type species, *Eoleperditia fabulites* (CONRAD, 1843) (cf. SWARTZ, 1949, p. 317). *E. skjesethi* sp.n. has, however, no subocular chevron mark, and may rather confidently be plased in *Eoleperditia* and not in *Leperditia* ROUAULT, 1851 or *Hermannina* KEGEL, 1933, at present not known from pre-Silurian strata (cf. SWARTZ, 1949). *E. skjesethi* differs from *E. fabulites* i.a. in having a larger number of spots in the adductor muscle scar, and in having a shorter hinge line. The muscle scars of *E. groenlandica* (TEICHERT, 1937) and *E. catheyensis* (KIRK, 1928) are not known, but they differ from *E. skjesethi* at least in their outline.

Occurrence: — Mjøsa Limestone at Nes—Hamar (Helgøya).

Other Ostracods.

Further research will probably show that the above described species are only a part of the ostracod faunas in the Middle Ordovician series in the Oslo Region. Besides the paleocop ostracods described above, there is also a *Bythocypris*-like smooth ostracod in the Upper Chasmops beds. The material of this is, however, not very good.

Remarks ön the Faunas.

The ostracod faunas of the Ogygiocaris Series (4a α) comprise forms which are not known from other areas. In the lower part of the Chasmops Series (4a β, 4b α—β) one species is also known from Sweden, where it is known to range throughout the Chasmops Series. The ostracods of the upper part of the Chasmops Series in the Mjøsa districts are all new, whereas those in the Oslo—Asker district, or at least closely related forms, are all known from Sweden and Esthonia. It is peculiar that they occur in the Lower Chasmops Series in Sweden and in the Kukruse (Kuckers) formation in Esthonia, which is believed to be of Lower Chasmops age.

Very little can thus be gained in the way of correlation from a comparision with other areas. With the exception of the Chasmops Series in Sweden and the Kukruse formation in Esthonia contemporaneous ostracod faunas have not been studied in detail in Europe (thus no ostracod faunas contemporaneous to those in the Ogygiocaris Series).

Several of the ostracod species described here, however, appear to be useful guide fossils within the Oslo Region.

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SPECIES	HORIZONS												LOCALITIES								
	Ogygiocaris Series				Chasmops Series														Mjøsa districts		
										Mjøsa districts											
	4 a α_1	4 a α_2	4 a α_3	4 a α_4	4 a β	4 b α	4 b β	4 b γ	4 b δ	Coel.	Cycl. = 4 b γ?	Mjøs. = 4 b δ?	Oslo—Asker	Gjerpen—Langesund	Eiker—Sandsvær	Modum	Ringerike	Hadeland	Toten	Nes-Hamar	Ringsaker
1. *Tallinnella trident* sp.n.	1	1	1												1	1					
2. » *tumida* sp.n.			2										2								
3. » *mjoesensis* sp.n.											3	3								3	
4. » *kiaeri* sp.n.											4								4	4	
5. *Ceratopsis*? sp.										5											5
6. » ? *stoermeri* sp.n		6													6						
7. *Piretopsis donsi* gen. & sp.n.					7								7								
8. *Steusloffia costata* (LINNARSSON)					8	8							8				8				
9. *Ullerella triplicata* sp.n.	9	9	9?												9	9					
10. » *ventroplicata* sp.n.				10									10				10				
11. » *ulli* (DONS)					11								11	11			11				
12. *holtedahli* (DONS)						12	12?						12				12				
13. *Bolbina variolaris* (BONNEMA)									13				13								
14. *Primitia suecica* (THORSLUND)								14					14								
15. *Platybolbina kapteyni* (BONNEMA)								15	15?				15								
16. *Euprimites*? *suecica* (THORSLUND)									16				16								
17. *Eoleperditia skjesethi* sp.n.												17								17	
18. *Bythocypris*? sp.									18				18								

STRATIGRAPHICAL AND GEOGRAPHICAL DISTRIBUTION OF MIDDLE ORDOVICIAN OSTRACODS OF THE OSLO REGION.

Coel. = Coelosphaeridium Beds. Cycl. = Cyclocrinus Beds. Mjøs. = Mjøsa Limestone.

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Manuscript received August 16. 1952.

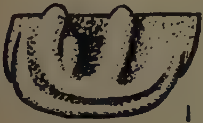
PLATES 1–5

PLATE 1.

Middle Ordovician ostracods from the Oslo Region. All enlarged approximately 10 times. Left valves only. (Reconstructions).

		Page
1.	*Tallinnella trident* sp.n.	37
2.	*Tallinnella tumida* sp.n.	38
3.	*Tallinnella mjoesensis* sp.n.	39
4.	*Tallinnella kiaeri* sp.n.	40
5.	*Ceratopsis ?* sp.	42
6.	*Ceratopsis ? stoermeri* sp.n.	42
7.	*Piretopsis donsi* gen. & sp.n.	43
8.	*Steusloffia costata* (Linnarsson)	43
9.	*Ullerella triplicata* sp.n.	46
10.	*Ullerella ventroplicata* sp.n.	47
11.	*Ullerella ulli* (Dons)	48
12.	*Ullerella holtedahli* sp.n.	48
13.	*Bolbina variolaris* (Bonnema)	49
14.	*Primitia suecica* (Thorslund)	49
15.	*Platybolbina kapteyni* (Bonnema)	50
16.	*Euprimites? suecica* (Thorslund)	51
17.	*Eoleperditia skjesethi sp.n*	52

PLATE 1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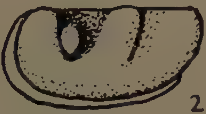
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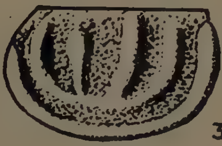
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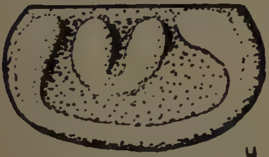
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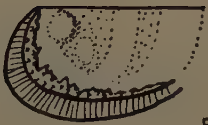
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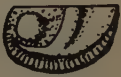
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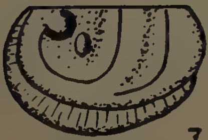
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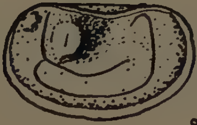
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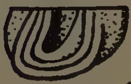
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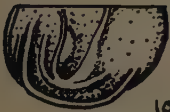
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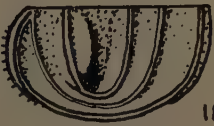
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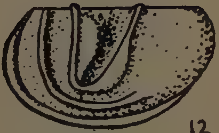
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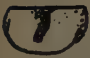
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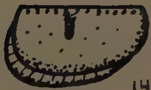
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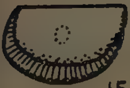
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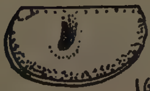
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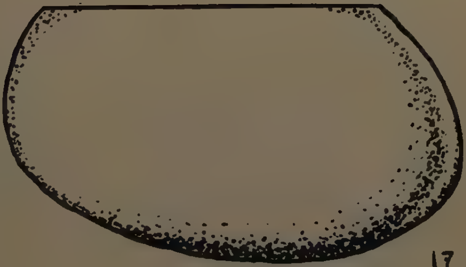
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PLATE 2.

The fossils have been whitened with ammonium chloride. The photographs are not retouched. Photographer: Miss B. Mauritz.

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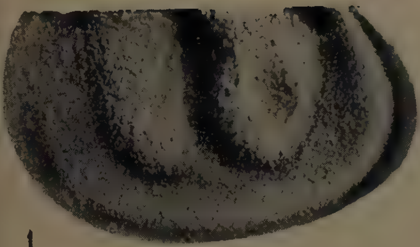
PLATE 2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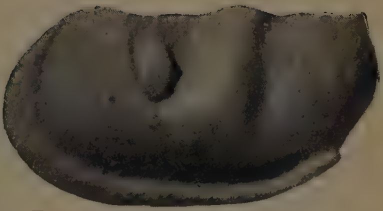
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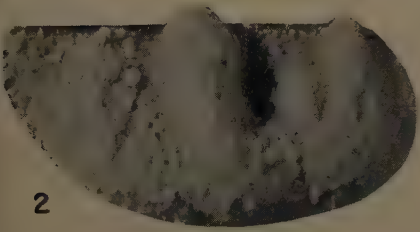
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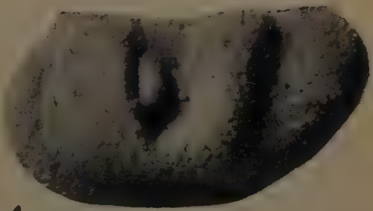
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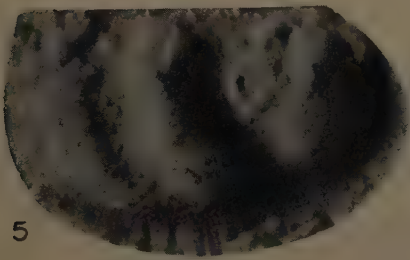
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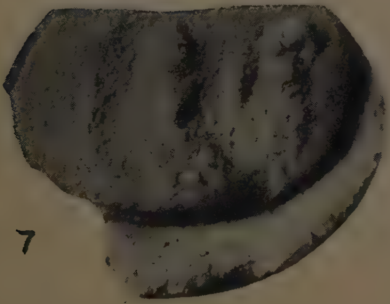
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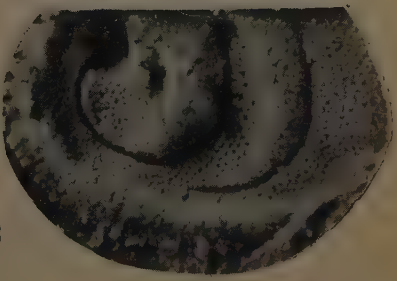
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PLATE 3.

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P.M.O. = Paleontological Museum, Oslo.

Tallinnella mjoesensis sp.n. (p. 39)

Cyclocrinus Beds. Furuberget, Nes—Hamar district.

Fig. 1. Right valve. P.M.O. 66422 (Holotype). C. 23 x

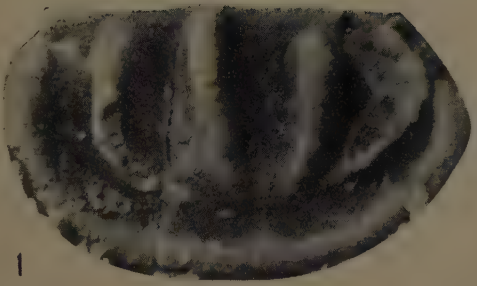
Figs. 2—4. Larval instars, right valves.

Fig. 2. P.M.O. 66433. C. 23 x

Fig. 3. P.M.O. 66432. C. 23 x

Fig. 4. P.M.O. 66431. C. 23 x

Fig. 5. Same as fig. 1 (Holotype). C. 40 x



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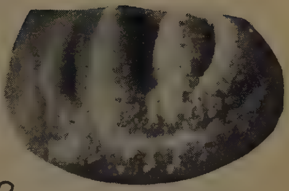

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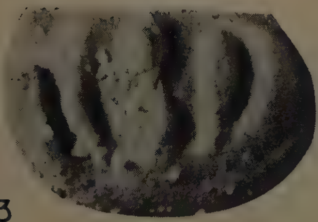

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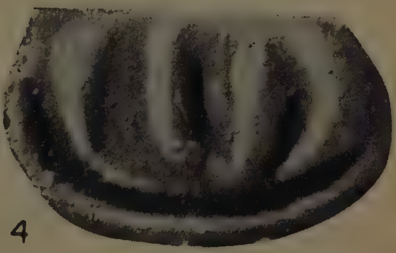

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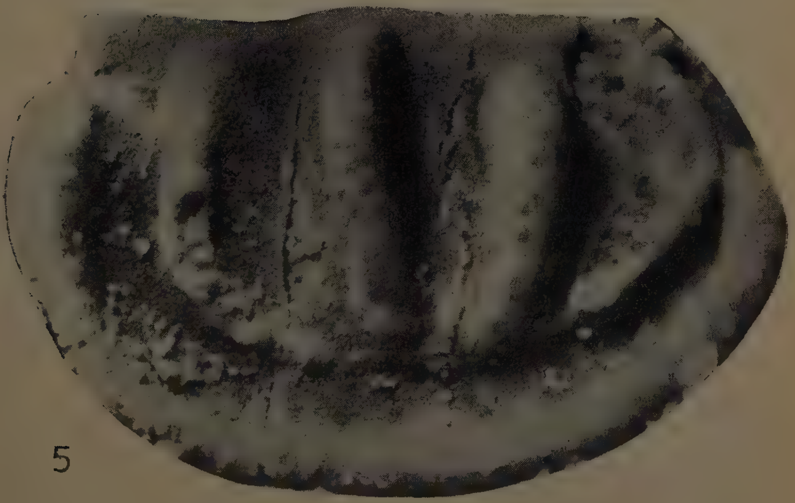

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PLATE 4.

The fossils have been whitened with ammonium chloride. The photographs are not retouched. Photographer: Miss B. Mauritz.

P.M.O. = Paleontological Museum, Oslo.

Tallinnella kiaeri sp.n. (p. 40).

Mjøsa Limestone, Furuberget, Nes—Hamar district.

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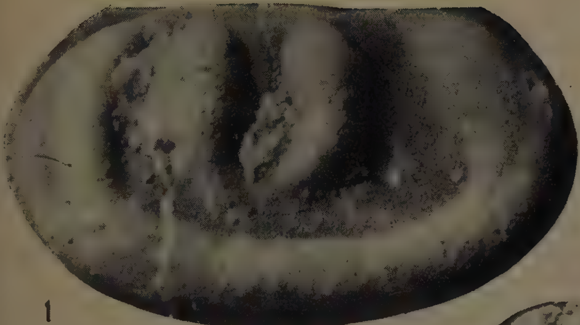
Fig. 2. Left valve. P.M.O. 66435. C. 13 x.
Fig. 3. Right valve. P.M.O. 66438. C. 20 x.
Fig. 4. Right valve. P.M.O. 66436. C. 13 x
Fig. 5. Right valve. P.M.O. 66437. C. 13 x.
Fig. 6. Left valve. P.M.O. 66434. C. 17 x.

PLATE 4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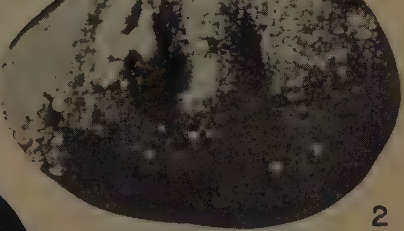
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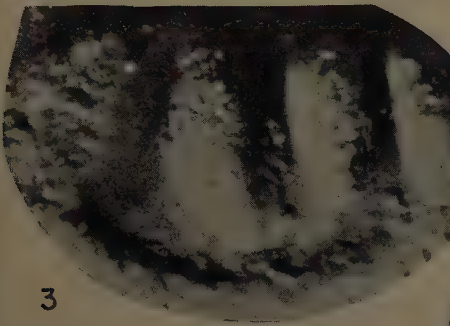
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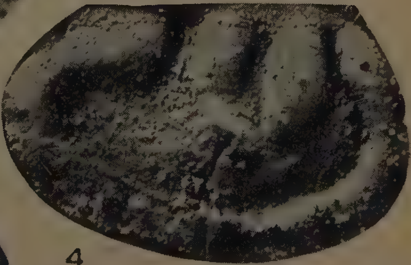
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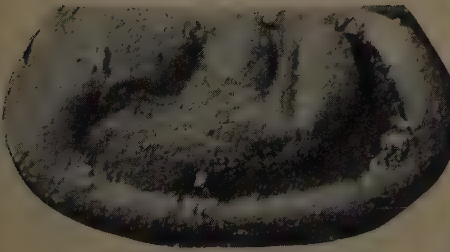
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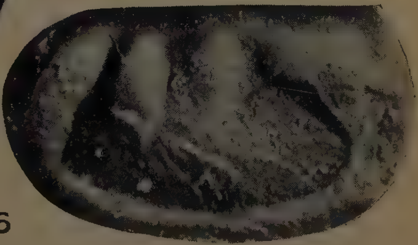
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PLATE 5.

The fossils have been whitened with ammonium chloride. The photographs are not retouched. Photographer: Miss B. Mauritz.

P.M.O. = Paleontological Museum, Oslo.

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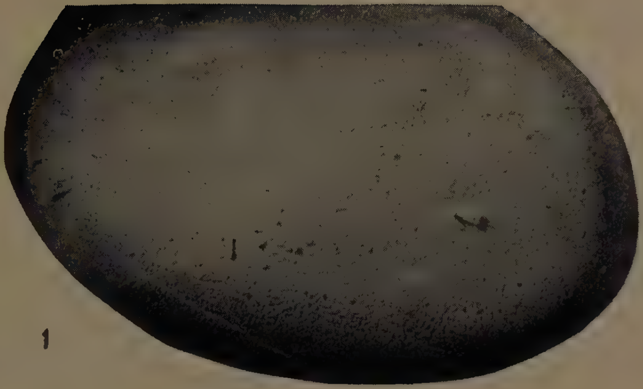
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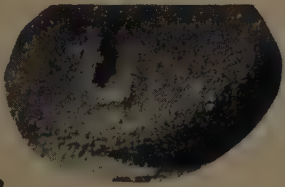
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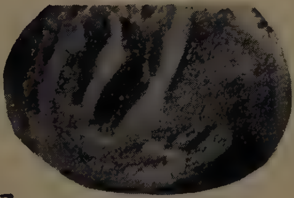
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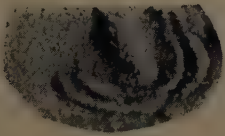
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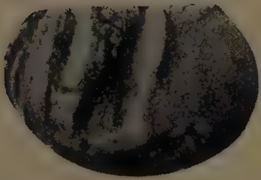
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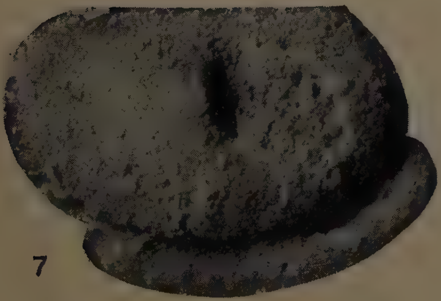
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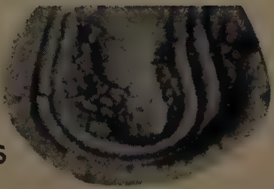
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A SELFCENTERING X-RAY CAMERA FOR POWDER DIAGRAMS

BY

HENRICH NEUMANN and JAN BARSTAD

X-ray powder diagrams are more and more widely used for mineral identification because of the usually unequivocal results and the conveniency of the method. Modern x-ray sets can easily be operated by an intelligent technician, and the sealed tubes, now on the market, are long lasting and efficient, reducing the time of exposure to a fraction of what it used to be a couple of decades ago. The most time consuming parts of the procedure are 1.) the centering of the pin, and 2.) the preparation of the pin. We have been able to reduce the time spent on both these operations to a minimum, and it may be worth while to publish our results.

Centering of the pin.

In most commercial cameras the pin is mounted on a holder more or less like a simplified goniometer head, and is centered by two or more screws; it takes normally a quarter of an hour or twenty minutes to get a good centering. For most people it will be a nearly irresistible temptation to try to save time by interrupting the centering before it is perfect, resulting in inferior films and inaccurate readings.

We have been able to avoid this operation altogether using a specimen holder with a hole in it which is centered once for all. When a rod is placed in a closefitting hole which is centered the rod itself must necessarily be centered. It appears from the drawing, Fig. 1., how the moving part of the camera is constructed, and a more detailed description should be unnecessary. The diameter of the centered hole is 0.12 millimeter and the pin is 0.10 mm thick.

With some experience it is not difficult to place the pin in the holder and it is only very rarely broken. When that does happen the holder is rejected and replaced by a new one.

Preparation of the pin.

The powder is mounted on a glass fiber of either lithium borate glass or ordinary glass giving identical results. A laboratory glass-rod is heated over a good Meker burner and by a fast movement drawn out to a fiber. Parts of the fiber which are not perfectly straight and measure 0.10 mm in diameter are rejected. In this way some hundred pins are easily made in half an hour. The pin is "wetted" by some adhesive or other, for example an extremely thin film of pure vaselin, and rolled in the powder to be examined. Unnecessary adhering powder is blown off, and the pin thus covered with an even film of powder is ready to be placed in the holder. The whole operation takes but a couple of minutes. A great advantage is that only very little material is needed; a grain just big enough to be seen by the naked eye is ample, and will give a perfect film.

Acknowledgement.

During the building of the camera a number of awkward mechanical problems were encountered, and skilfully solved by Mr. Jahrl Andersen, the mechanic of The Geological Museum, Oslo, whose cooperation and expert workmanship is hereby duly acknowledged.

University of Oslo, Geologisk Museum, September 1952.

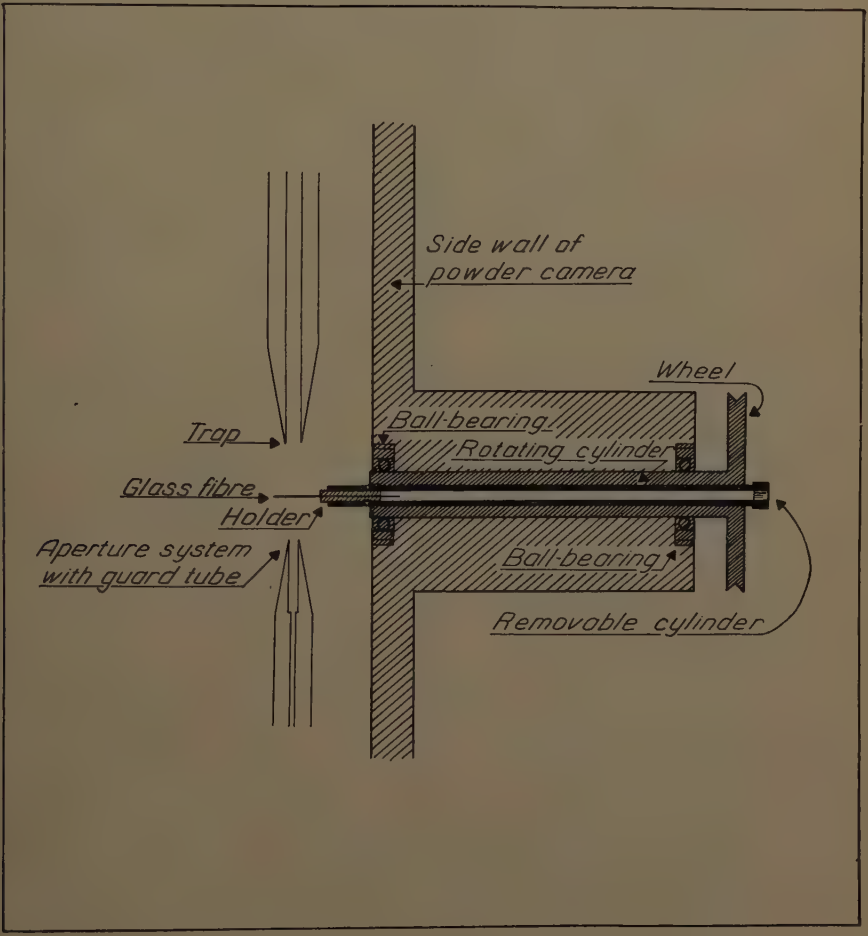

Fig. 1. Moving parts of selfcentering x-ray camera.

NORSK GEOLOGISK TIDSSKRIFT 32

THE FeS-ZnS SYSTEM
A GEOLOGICAL THERMOMETER

BY

GUNNAR KULLERUD
(Mineralogisk-Geologisk Museum, Oslo, Norway).

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Abstract. In the first part of this paper, the FeS — ZnS equilibrium diagram is presented. Laboratory equipment and methods used for the establishment of the diagram are described. The end members of the system as well as the α and β mix-crystals are studied. The ZnS α-β inversion mechanism has received special attention, and the relationship between inversion temperature and mix-crystal composition is established. The increase in the ZnS lattice dimensions due to the replacement of Zn by Fe is explained by assuming that the mix-crystal bonds acquire a more ionic character as their FeS content increases. Thermodynamic calculations are applied for determinations of the equilibrium curves below 400° C.

The influence of pressure upon the FeS solubility in the ZnS lattice has been calculated, and the results in part checked by means of laboratory experiments.

The FeS activity was investigated and was found to decrease very little from stoichiometric FeS to FeS_2. However, it drops sharply, when S is added beyond the amount indicated by the FeS_2 formula. Studies were conducted of the unmixing of FeS from mix-crystals. It was also found that small amounts of Mn and Cd have little or no influence on the solubility of FeS in ZnS.

In the second part, chemical, spectrographic and x-ray methods for determination of the composition of sphalerite ore specimens are described. Results of the analyses together with the temperatures deduced from the composition of the samples are presented in tabular form. Some of the localities from which specimens were obtained are described with a view to ascertaining the pressure conditions prevailing during ore deposition. Discussions whether the mix-crystals contain as much FeS as prescribed by the (P,T) conditions of ore formation, are often of little value unless coupled with field work and microscopic studies. While FeS in many cases seems to have been present in ample quantities to satisfy the equilibrium requirements, Mn, Cd, and the trace elements never occurred in sufficient amounts to saturate the ZnS lattice. The only reliable index of temperature of formation is FeS. The distribution of the other elements is not necessarily related to this temperature. Possible applications of the FeS — ZnS system to geological problems are outlined.

PART I.

Introduction.

The temperature-pressure conditions existing in rocks and ore bodies at the time of their formation, have been discussed in a great number of publications. Several methods have been applied for determinations of geologic temperatures. In some instances geological temperatures have been directly measured, for example in lavas, but in most cases the temperatures have been deduced from ascertained facts regarding the effect of heat on various physical properties of minerals. Of such methods may be mentioned high-low inversion of quartz, liquid inclusion methods, etc.

As has been emphasized by a number of authors, the concentrations of elements in mix-crystals under certain conditions may serve as our most reliable geological thermometers, provided enough thermodynamic data of the minerals are known. The general principles underlying such thermometers are clear, but in no case is enough data available for an accurate and absolute thermometer to be established among the several systems in which mix-crystals occur. Among the ore minerals the importance of mix-crystals as a guide to geologic thermometry has been discussed, but so far nobody has thoroughly investigated any of the mix-crystal series in the laboratory.

The FeS-ZnS Solid Solutions.

Many sulfide minerals can enter into solid solution with one another, and in some instances the solid solutions persist as homogeneous minerals, stable at normal temperatures.

It has long been known that ZnS has the ability to dissolve appreciable amounts of FeS. In nature pure, colorless ZnS is extremely rare. Most sphalerites contain iron and traces of other elements and are no longer colorless, but yellowish to brown, sometimes almost black depending upon the amount of iron dissolved. Provided sufficient iron sulfide was present during the formation of the mix-crystals, the ZnS would dissolve an amount of iron determined by the temperature and pressure conditions existing during the formation of the sulfides.

In numerous ore deposits pyrrhotite occurs intimately connected with sphalerite (marmatite). From geological studies it has been shown that an abundance of iron was often present during the entire process of ore formation. Therefore, in such cases one may safely assume that equilibrium conditions existed between free pyrrhotite and the FeS dissolved in the sphalerite lattice at the time of sphalerite formation. In any sample of sphalerite fulfilling the above-mentioned conditions, the amount of FeS dissolved is related to the temperature of formation of mix-crystals. Hence, the temperature of ore-deposition, disregarding pressure, may be determined from the amount of FeS dissolved in the sphalerite lattice using the FeS-ZnS diagram (fig. 1). The modifying influence of pressure upon the equilibrium is demonstrated in fig. 9.

The FeS-ZnS System.

Very little work has previously been done on the FeS-ZnS system. Friedrich[1] investigated the melting curves and found a eutectic to exist in the system for 5 % ZnS at 1170° C. Later Allen and Crenshaw[2] investigated the effect of iron on the inversion point of

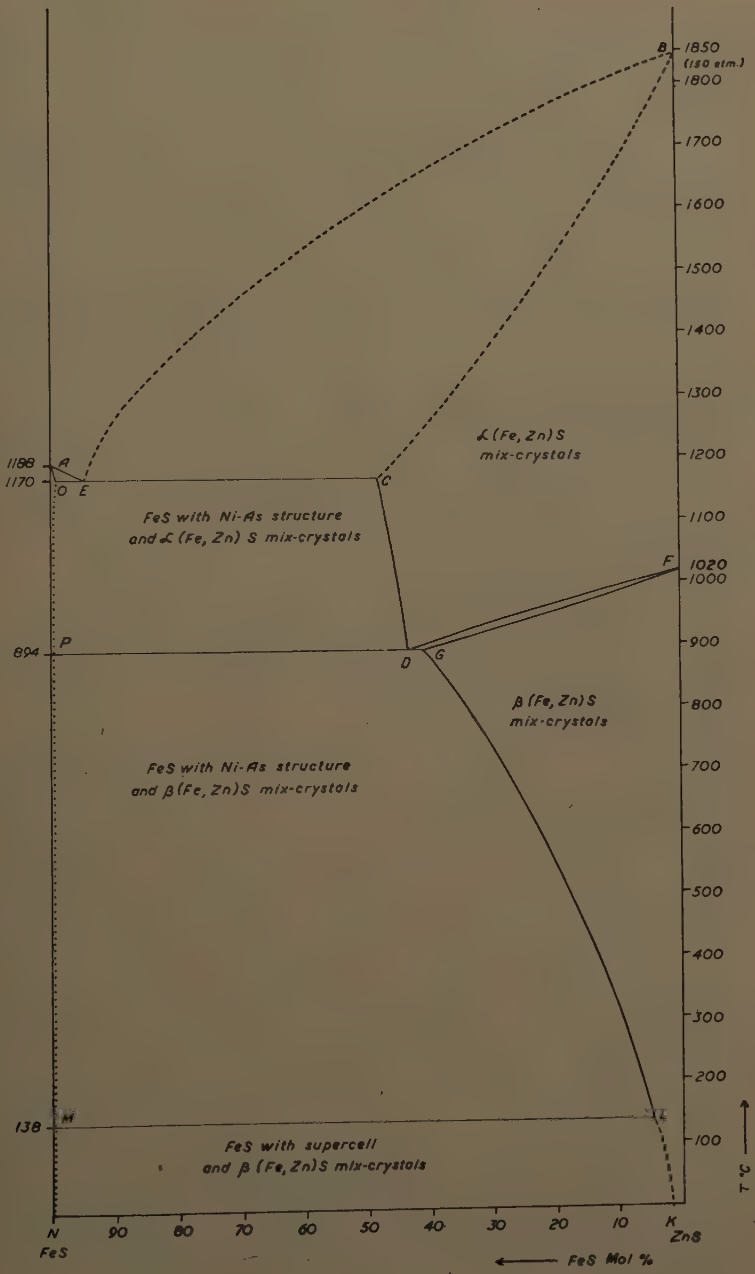

Fig. 1. The FeS-ZnS equilibrium diagram.

sphalerite. They showed that the sphalerite-wurtzite inversion temperature decreases with increasing amount of iron dissolved in the ZnS lattice.

The complete FeS-ZnS phase diagram is shown in figure 1. This diagram was drawn as a result of more than 250 runs with mixtures of FeS and ZnS in varying proportions. The time needed for equilibrium to be established in the system varied from a few hours at 1000° C to nine months at 500° C.

The liquidus curves AE and part of EB were established by FRIEDRICH, his values as well as values for parts of these curves given in the "International Critical Tables" have been used in Fig. 1. Below the eutectic temperature, 1170° C, there are two series of solid solutions. The solubility of ZnS in hexagonal FeS, however, is, even at 1160° C, less than ½ % by weight and is of no practical importance for the later application of the diagram. This solid solution series is suggested by a stipled line (OPMN). The solubility of FeS in α and β ZnS is appreciable as seen from the curves CD and GLK.

Zinc sulfide occurs in two enantiotropic modifications. β ZnS (sphalerite) with a cubic face-centered lattice, being the low temperature modification, converts at 1020° C into α ZnS (wurtzite) with a hexagonal, close-packed lattice.

The FG curve was partly determined by ALLEN and CRENSHAW as mentioned above. For their studies these investigators, however, used sphalerite ores containing impurities of Mn and Cd, each of which depresses the inversion point even more than iron does. Therefore, their results were somewhat lower than the ones obtained by the present author in his runs with pure sulfides of iron and zinc.

Pure FeS melts at about 1188° C under atmospheric pressure. The pure α ZnS on the other hand does not melt at normal pressure, but sublimates at 1185° C. It is reported to melt at 1850° C under 150 atm. pressure.[1] Therefore the upper right part of the diagram (fig. 1) falls slightly outside the isobaric section represented by the rest of the diagram.

From fig. 1 it is seen that above the curves AE and EB there exists one melt. In the fields AOE and BEC liquid and solid coexist, the composition of the liquid being determined by AE and BE and the composition of the solid being determined by the solidus curves AO and BC. In the field AON one finds mix-crystals formed by solution of ZnS in FeS. In the field BCDF α (Fe,Zn) S mix-crystals occur, and in the FGK area β (Fe,Zn) S mix-crystals are stable. The separating FGD field contains both α (Fe,Zn) S and β (Fe,Zn) S mix-crystals.

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[1] International Critical Tables.

The End Components of the FeS-ZnS System.

Before going any further in the discussion of the presented diagram and the description of laboratory equipment and methods, it may be appropriate to look into the properties of pyrrhotite as well as the two enantiotropic modifications of zinc sulfide.

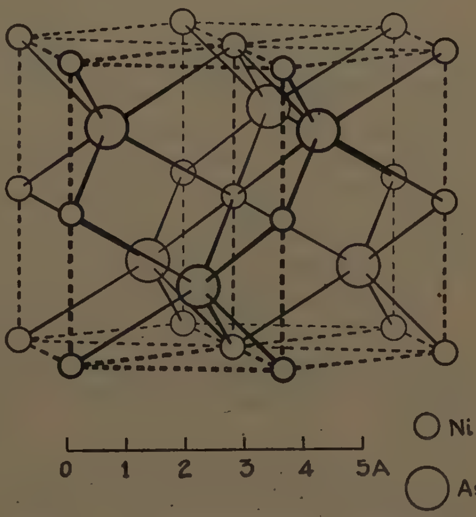

Fig. 2. The NiAs structure. Hexagonal FeS is obtained when Ni is replaced by Fe and As by S. (Strukturbericht).

1. *Pyrrhotite.*

The crystal structure of pyrrhotite is of the Ni-As type.[1] Crystals of this type are often associated with a super-cell which makes the understanding of these crystals hard.

According to the literature, ferromagnetic pyrrhotite exists in the composition range $Fe_{48}S_{52}$ to $Fe_{466}S_{534}$ and has the nickel-arsenide type structure (see fig. 2). Paramagnetic pyrrhotite is confined to the composition range $Fe_{50}S_{50}$ to $Fe_{48}S_{52}$ and has two kinds of structures: nickel-arsenide type in the high sulfur half of the field and a super-cell in the region near ideal composition FeS (the border being about $Fe_{49}S_{51}$). FeS at room temperature shows the super-cell which has a hexagonal or trigonal symmetry, and contains 12 molecules; and which in volume is six times as large as the normal cell. The a_o ($5.946A \approx 3.453 . \sqrt{3}$) of this lattice is the long diagonal of a_o (3.453A) for the simple Ni-As cell; its c_o (11.720A) is approximately twice as great as that of the Ni-As cell (c_o 5.860A). (FeS_x: $a_o = 3.43A$, $c_o = 5.68A$). Whether this so called super-cell has the characteristics of a super-lattice has to the author's knowledge not so far been investigated.

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[1] Whether natural pyrrhotite has true hexagonal crystallographic properties or is of a pseudo-hexagonal character has been discussed by BUERGER[3]. After having investigated single crystals of pyrrhotite from two localities, Buerger concluded that pyrrhotite either belongs to a low hexagonal or to an orthorhombic or even may be monoclinic symmetry.

On heating, the super-cell is transformed to the Ni-As arrangement. The transition point depends upon the composition of the pyrrhotite. Stoichiometric FeS inverts at 138° C while pyrrhotite of composition $FeS_{1.09}$ passes from the super-lattice to the simple Ni-As grouping at the lower temperature of 74° C. The Ni-As structure in turn suffers another transition at 325° C involving a discontinuous change of spacing but not of atomic arrangement. These transitions have been studied by HARALDSEN[4, 5, 6, 7, 8, 9] who presented a diagram showing the transition points and variations in lattice parameters for FeS as related to the temperature.

Variation in the sulfur content of the pyrrhotite series of which troilite (FeS) is considered an end member has been shown by ALLEN, CRENSHAW, JOHNSEN and LARSEN[10] to be as much as 6.5 %, the equivalent formula being $Fe_1S_{1.18}$. The high sulfur ratio has been shown (LAVES[11], HÄGG[12], HÄGG and SUCHSDORFF[13]) to be due not an excess of sulfur, but to iron atoms missing from their structural positions. Recent work by JENSEN[14] shows that if this explanation holds at high temperatures, the lattice has a maximum stability when there are a certain number of vacant iron positions, corresponding to a composition of $FeS_{1.08}$ or $Fe_{12}S_{13}$.

According to JENSEN the solubility of iron in FeS is negligible, though there is a possibility that it may be greater at high temperatures.

A review of the literature by JENSEN shows that the limit of solubility of S in FeS at low temperatures is about 39.1 % (as compared with 36.5 % S in FeS). This limit increases with the increase in temperature to about 40.7 % at 677° C.

The first x-ray examination of a mineral belonging to the nickel-arsenide group was undertaken by AMINOFF[15]. In this structure each atom has six nearest neighbors of the other sort, but the arrangement of the neighboring atom is different for the two kinds of atoms. Thus in pyrrhotite a sulfur atom is surrounded by six Fe atoms at the apices of a trigonal prism. The immediate neighbors of an Fe atom are six S atoms arranged octahedrally, but there are also two Fe atoms sufficiently near to be considered bonded to the first Fe atom (Wells[16]).

The minerals with the Ni-As structure often have a metallic character (Bragg[17]) besides the mentioned variation in composition. In the more metallic systems with this structure the eight neighboring atoms are approximately equidistant from the metal atom. As a general rule the closer the atoms approach equidistant positions the more metallic are the bonds. (For instance for CoSb the Co-Sb distance is 2.58A and the Co-Co distance 2.60A. The corresponding distances for FeS are 2.45A and 2.89A. CoSb which has equidistant atoms is much more metallic than FeS whose atoms are far from being equidistant).

2. *The β ZnS* (*Sphalerite*).

In sphalerite the zinc atoms are arranged on a face-centered cubic lattice. The sulfur atoms, arranged on a similar lattice, are so located that each is between four zinc atoms. Figure 3 shows only the four sulfur atoms included within the unit cell, but it will be clear that if the pattern is continued they will build up a lattice like

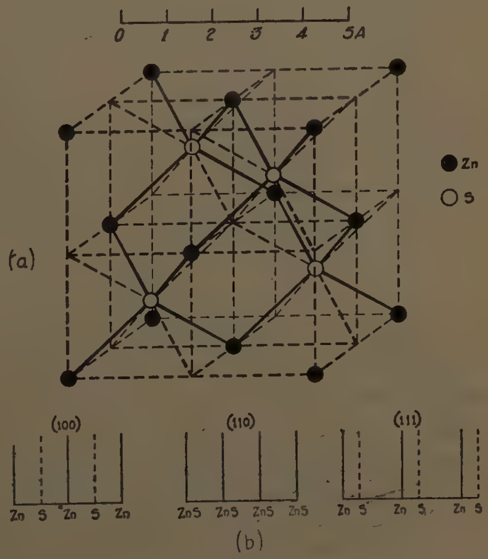

Fig. 3. a. The structure of sphalerite, ZnS. b. The arrangement of planes parallel to (100), (110), and (111), the last illustrating the polar nature of the [111] axis. (Bragg[17]).

that of the zinc atoms. The arrangement is like that of diamond, if one replaces both Zn and S with C. Whereas diamond is holohedral, the distinction between Zn and S causes the present structure to be hemihedral.

Figure 3 shows that the sulfur atoms are surrounded by four zinc atoms at the corners of tetrahedra which are always similarly oriented. Hence the whole structure has the symmetry of a tetrahedron, not that of a cube. This is seen in an alternative way by the arrangement of the [111] planes shown in figure 3b. These planes are alternately Zn and S in pairs, and therefore the direction perpendicular to them ([111] axis) is polar. Opposite faces [111] and $[\dot{1}\dot{1}\dot{1}]$ are not identical in character. The characteristics are in accord with the symmetry of sphalerite crystals displayed on etching.

The number of Zn and S atoms (Z) in the unit cell is Z = 4. The length (a_w) of the unit cell has been determined by several workers. The literature gives values varying from $a_w = 5.39$ to 5.42 A. In crystals with the diamond, sphalerite, or wurtzite arrangement each atom is surrounded tetrahedrally by four other atoms. If the atoms are those of fourth-column elements or of two elements symmetrically arranged relative to the fourth column, the number of valence electrons is right to permit the formation of a tetrahedral covalent bond between each atom and its four neighbors. In ZnS an extremely covalent structure places the formal charges 2^- on zinc and 2^+ on sulfur. It is probable that the bonds have enough ionic character in this crystal and others of similar structure to make the actual charge of the atoms much smaller.[18] Wells[16] also assumes that the ZnS bonds must have appreciable ionic character. The ZnS bonds it may be concluded are intermediate between ionic and covalent bonds.

3. *The a ZnS (Wurtzite).*

This high temperature form of zinc sulfide is hexagonal, and its structure is related to that of β ZnS in a simple way.[19] The relation resembles that between cubic and hexagonal closest packing. Zinc atoms in blende are on a face-centered cubic lattice, whereas in wurtzite they are nearly in the positions of hexagonal closest packing. This, as pointed out by Bragg, is the alternative of close packing and produces just as dense a structure as the face centered cubic packing provided the hexagonal axial ratio c/a is 1.633 (which is twice the ratio of the height of a regular tetrahedron to its edge). The wurtzite tetrahedra are practically regular. (That means c/a is almost exactly 1.633. The same is true of CdS). In both structures the sulfur atoms lie between four zinc atoms. The wurtzite structure is shown in figure 4.

If the structure of blende is pictured standing on a tetrahedron face, the relation between blende and wurtzite may be seen without difficulty. The zincblende structure may be produced from the wurtzite structure by a rotation of 60 degrees of alternate pairs of [0001] atomic layers about the threefold axis. The hexagonal axis of wurtzite is polar. The sheets of atoms parallel to the basal plane all have Zn above and S below in each pair (CdS has a similar structure). The distances between the atoms are practically identical with those in zincblende. Shôji[20] showed that the transformation planes may be the [111] for zincblende and the [0001] for wurtzite. The hexagonal a and c value calculated when using the zincblende [111] as basal plane were only about 0.7 % smaller than the values calculated from x-ray diagrams.

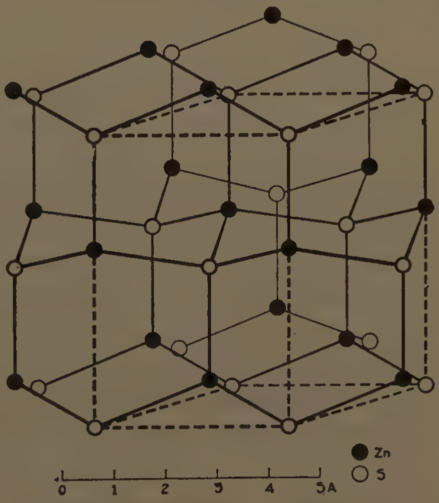

Fig. 4. The structure of wurtzite, ZnS. (Strukturbericht).

Equipment for the Runs.

The essential parts of the furnaces used are the core (1¼″—1½″ diameter) of alundum with nichrome wire wound around and cemented to the core, and insulating fire bricks (cut to fit around the core and inside a cylindric mantel 10″ diameter). These furnaces could be run at 900—1000° C for long periods and as high as 1100° C for shorter periods of time. The energy input was controlled with "temcometers". Chromel-Alumel thermocouples were used between 200° and 1000° C. For the same temperature interval a "Brown Electronik" recorder gave temperature readings every 1½ minutes on a continuously moving chart. Platinum-rhodium thermocouples in connection with a Rubicon potentiometer were used for readings of temperatures higher than 1000° C.

With this equipment it was possible to control and record the temperatures within ± 5° C of the real values.

The Bombs.

Because of the high vapor pressure of sulfur it was necessary to let the sulfide formations take place in closed bombs made of a material that would not react with the enclosed substance. As material for the bombs, transparent quartz glass tubing was chosen.

The bombs had to be evacuated in order to prevent oxides forming and thus stagnate or completely prevent the sulfide formation.

While iron sulfide forms easily from iron and sulfur at a temperature as low as 500° C, it was expected to be necessary to heat the mechanical compound of zinc and sulfur at more elevated temperatures in order to obtain a complete zinc sulfide reaction.

The bombs therefore had to be made of a tubing with thick enough walls to withstand the vapor pressures of sulfur and zinc at such temperatures. It was believed that it would not be necessary to exceed 850° C in order to have the zinc sulfide formation take place. The vapor pressure of sulfur is about 40 atm. at 850° C, while the vapor pressure of zinc at the same temperature is less than 1 atm. (boiling point for Zn is 907° C).

Quartz glass tubing with an internal diameter of 6 mm and a wall thickness of 3 mm was found to withstand this pressure and therefore was used for the runs. For later runs where ready made sulfides were mixed together, tubing with thinner walls was used, and in cases where a very small amount of substance was needed, the bombs were made of tubing with 2 mm inside diameter and 1 mm thick walls.

The procedure of making these bombs was as follows: The quartz glass tubing was fire polished at one end over an acetylene flame so as to form a closed, rounded end. The substance was then inserted in the tube and the tube again heated about $1\frac{1}{2}$ to 2 inches from the end and drawn out into a capillary. The tube was then connected with a Cenco Hyvac pump which gave a vacuum of about 0.3 Micron Hg. The air in the system had to be pumped out slowly and carefully so that no substance would be drawn out with it. This was done by controlling the rate of air flow with a partially closed stopcock. After a suitable vacuum was obtained the capillary was heated again and the bomb sealed off.

The Formation of FeS and ZnS.

Pyrrhotite, of almost stoichiometric FeS composition, (very slight excess of sulfur) was made from analytically pure iron and sulfur, in a bomb of the type previously described. The bomb was heated at 540° C for 48 hours and then quenched. The resulting hexagonal pyrrhotite crystals were about 1/30 mm long, paramagnetic and of a medium dark grey color. X-ray powder diagrams of this substance showed the Ni-As arrangement.

Sphalerite was made from analytically pure zinc and sulfur mixed together in the proportion given by the ZnS formula. The formation of this sulfide did not take place as easily as that of pyrrhotite. A very high vacuum in the bomb and heating for 7—8 days at a minimum of 870° C proved to be necessary in order to obtain a complete reaction. In the first few runs the required vacuum was not obtained, consequently ZnO layers were formed on the Zn-crystals preventing ZnS to form.[1] The sphalerite formed was of an almost pure white color and occurred in cubic crystals the size of which was approximately 1/25 mm. Some of the sphalerite was heated at 1050° C for 8 hours in a closed bomb. The resulting hexagonal wurtzite crystals were of a slightly yellow color.

The Mix-crystals.

Solid solutions were found to form from mixes of the pure sulfides. The mix-crystals readily formed at temperatures above 850° C.

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[1] An interesting study of the formation of ZnO films on Zn crystals has recently been done by RAETHER[21]. The oxide film formed on exposure of metallic zinc to the air, was monocrystalline and parallel to the orientation of the Zn crystal lattice. The thickness of the film was 20—40 A at room temperature, and increased at elevated temperatures.

Below this temperature mix-crystals formed only after prolonged heating. In order to establish the equilibrium curves of the FeS-ZnS diagram, the composition of the mix-crystals, formed at given temperatures, had to be determined. Therefore the optical properties as well as the specific gravities and unit cell dimensions of mix-crystals of known compositions, were studied thoroughly. The refractive index was found to increase with increasing iron content, this being in agreement with results obtained by ALLEN and CRENSHAW[2] who determined the refractive indices for sphalerite ores with varying amounts of iron. The specific gravity was found to decrease with increasing iron content, a fact also observed previously by the above mentioned workers. Further, an increase in the unit cell dimensions with increasing iron content was observed by means of x-ray powder diagrams. X-ray studies of the variations in the cell dimensions of sphalerites with varying amounts of iron have previously been undertaken by RINNE,[22] GOLDSCHMIDT,[23] BRAEKKEN,[24] and CHUDOBA.[25] The former three authors, with the x-ray equipment available at that time, found no expansion in the lattice caused by the presence of iron. CHUDOBA, who calculated the spacings from x-ray powder diagrams of four sphalerite ores, found different unit cell length values for all four.[1] At the time the present work was started, little information, therefore, existed concerning the influence of iron alone on the spacing of sphalerite.

After comparing the three above mentioned methods for determination of the composition of (Fe, Zn) S mix-crystals, the x-ray method was considered the best suited. [With the refractive index method the iron content could be determined within 1 %. Pure sphalerite has a refractive index n = 2.36 and the mix-crystals with maximum iron content go up to n = 2.525. As no liquid goes that high, compounds of sulfur and selenium were used for the determinations of these high refractive indices. The sulfur-selenium compounds are not as convenient to use as liquids and are not very desirable for exact determinations.

The specific gravity was determined by the BILTZS'[26] pycnometer method and an accuracy of ± 0.003 was obtained. The specific gravity is d = 4.110 for pure sphalerite and 3.935 for (Fe, Zn) S with 36½ % FeS by weight. This method gives fairly accurate determinations of the Fe content of the mix-crystals (about ½ %), but is slow. The x-ray powder diagram method gave by far the best results when a suitable internal standard was used.]

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[1] it may here be mentioned that two of the specimens, used by CHUDOBA for his experiments, also contained a substantial amount of manganese and some cadmium, which both influence the spacing.

The X-ray Work.

The x-ray powder diagrams were taken in an improved Debye-Scherrer cylindrical camera (diameter D = 114.6 mm) designed by the North American Philips Co., Inc. The various factors entering the design of this camera and interrelated in their influence on the properties of the patterns are considered in a paper by W. PARRISH and E. CISNEY.[27] This camera has a light-tight cover so that it is not necessary to make the exposures in a dark room or to cover the film with black paper. The primary x-ray beam passes through two conical tubes between which the thin rod-shaped specimen is mounted. After placing the specimen in a slot on the axis of the camera, accurate centering is done by means of a threaded "pusher" perpendicular to the axis. During exposures a cap containing a fluorescent screen is placed on the end of the exit tube. On this screen the primary beam and the shadow of the specimen may be observed. This fluorescent screen is covered on the inside with black paper making the camera light-tight, and on the outside with lead glass absorbing the remainder of the primary beam. The film is placed in the camera according to Straumanis method: the film takes up nearly 360° and two holes of 9 mm diameter are punched in it where the tubes are to be inserted. The film is expanded against the inner wall of the camera by means of a movable finger which pushes one end of the film, the other end being fixed by a stop. While the exposure is being made the specimen is rotated continuously around its axis by a pulley attached to a small electric motor.

The sample was mixed with Duco Household Cement and rolled out between two glass plates to form a cylindrical rod. Much care was taken in giving this rod a true roundness and an even thickness.

In order to investigate what influence the rod thickness has on the sharpness of the lines of the powder diagrams, several rods of different thicknesses were made from pure sphalerite. It was found increasingly difficult to obtain sharp lines as the rod thickness d, exceeded 0.4 mm. Very good sharpness was achieved for d = 0.35 mm. The sharpness obtained for this thickness did not differ noticeably from that obtained for d = 0.25 mm. For this reason the rods were made approximately 0.30 mm thick; a tolerance of 0.05 mm was allowed

The centering of the rods was done with much care to obtain the best possible accuracy.

The films were measured by means of a recorder (Picker) giving readings as exact as ± 0.05 mm.

It was calculated that for the smallest Θ used (about 58°) the biggest possible error in spacing, caused by an uncertainty of ± 0.05 mm in reading was ± 0.00003A. In order to compensate for film shrinkage, variations (d = 0.25—0.35 mm) in rod thickness and a series of systematical errors, about 30 % rock salt was mixed in as internal standard with the samples.

Wyckoff[28] gives for NaCl, $a_o = 5.62869$A. Fe radiation with Mn filter was used for the powder diagrams. (Cu radiation with Ni filter was tried and found excellent for pure ZnS, but when more than 5 % Fe is present, the background of the film gets almost as dark as the reflections and accurate readings are increasingly harder to obtain.) The wave lengths used were:

$$\text{Fe K}\alpha = 1.9373 \text{ A}$$
$$\text{Fe K}\alpha_1 = 1.93597 \text{ A}$$
$$\text{Fe K}\alpha_2 = 1.93991 \text{ A}$$

The exposure time varied from 20—25 hours. The NaCl [440] and [422] lines were used as standard lines and their Θ values were computed from the a_o value given above and the wave lengths. The β ZnS lines used for the calculations were the [333] and [422] lines. These reflections are located between the above mentioned NaCl lines.

Lattice Dimensions of Pure β ZnS and the β (Fe, Zn) S Mix-crystals.

Several x-ray powder diagrams were made of the pure β ZnS. The calculated unit cell lengths varied from $a_o = 5.39837$ to $a_o = 5.39861$ A, the mean value being $a_o = 5.3985$ A. Out of eight determinations six were inside the limit of error of $5.398_{5\pm1}$A.

The a_o values of the mix-crystals were determined from a smaller number of powder diagrams (each film was measured twice). Therefore the limit of error is bigger than that given for pure β ZnS. However, when only a couple of x-ray diagrams were made of a given mix-crystal composition and the a_o values were calculated, the discrepancy in these values never exceeded $\pm$ 0.0002A. This value will be given as the limit of error in the a_o values for the β mix-crystals.

Table I gives the final results of experiments carried out to determine the relationship between mix-crystal composition and unit cell length. In the last two vertical columns are given the molecular volumes and specific gravities calculated from the a_o values. (The N value used for these calculations is $6.02 \cdot 10^{23}$).

TABLE I.

Number of runs.	FeS content of β mix-crystals		Unit cell length a_o in A	Calculated molecular volume	Calculated specific gravity
	wt.%	mol.%			
4	0	0	$5.398_{5\pm1}$	39.3332	4.1150
6	5	6.69	$5.400_{0\pm2}$	39.3594	4.0921
6	10	11.98	$5.401_{8\pm2}$	39.4054	4.0675
6	15	17.80	$5.404_{3\pm2}$	39.4601	4.0417
6	20	23.48	$5.406_{6\pm2}$	39.5105	4.0163
4	30	34.47	$5.410_{9\pm2}$	39.6049	3.9669
5	36.5	41.36	$5.413_{4\pm2}$	39.6598	3.9359

The results of the four runs listed in the upper column of table I have been mentioned already (see above).

In the next five horizontal columns the results from a total of twenty-eight runs are presented. The details of these runs are found in table II.

It is believed that the data given in table II are presented so clearly that individual discussion of each run is not required. In the first vertical column is given the amount of FeS mechanically mixed in with β ZnS. These compounds were then heated for a period of time given in column two, at a temperature given in column three. The a_o values of the resulting β mix-crystals are presented in column four. In columns five, six and seven are shown the results of a number

TABLE II.

Composition: FeS mixed with ZnS	Duration of runs	T°C	Unit cell length a_o in A	Duration of runs	T°C	Unit cell length a_o in A
5 wt.% (6.69 mol.%)	1 week	850	5.4001	2 weeks	750	inhomog. substance
	2 weeks	850	5.4000	3 weeks	750	5.3999
	3 weeks	850	5.4001	4 weeks	750	5.4001
10 wt.% (11.98 mol.%)	1 week	860	5.4015	2 weeks	750	inhomog. substance
	2 weeks	860	5.4018	4 weeks	750	5.4017
	3 weeks	860	5.4018	6 weeks	750	5.4017
15 wt.% (17.80 mol.%)	1 week	850	5.4041	2 weeks	750	inhomog. substance
	2 weeks	850	5.4044	4 weeks	750	5.4041
	3 weeks	850	5.4043	6 weeks	750	5.4043
20 wt.% (23.48 mol.%)	1 week	850	5.4064	3 weeks	750	5.4063
	2 weeks	850	5.4065	4 weeks	750	5.4065
	3 weeks	850	5.4066	6 weeks	750	5.4065
30 wt.% (34.47 mol.%)	4 days	890	5.4105			
	10 days	890	5.4108			
	3 weeks	890	5.4109			
	4 weeks	890	5.4109			

of runs carried out with compounds of the same compositions as before but at the lower temperature of 750° C.

It appears (table II) that the time needed for the formation of homogeneous mix-crystals at 750° C is much longer than it takes for such mix-crystals to form in the 850—890° C temperature range. In no instance homogeneous products were formed in the course of two weeks at 750° C, while at 850° C one week was sufficient, and at 890° C only four days were required.

In the last horizontal column of table I is given the maximum amount of FeS that can be dissolved in the β ZnS lattice (at 894° C where the β mix-crystals containing a maximum amount of FeS invert to the α modification) and the corresponding a_o value. These data will be further treated when the α-β inversion curves for the mix-crystals are studied.

Runs conducted at 890° C with compounds consisting of 40 wt.% (45.0 mol.%) FeS mixed with β ZnS produced mix-crystals with a_o = $5.413_{1\pm2}$A which is bigger than the a_o value obtained from runs containing 30 wt.% (34.47 mol.%) FeS. However, this increase is smaller than that obtained when the FeS content was increased from 20 wt.% (23.48 mol.%) to 30 wt.% (34.47 mol.%).

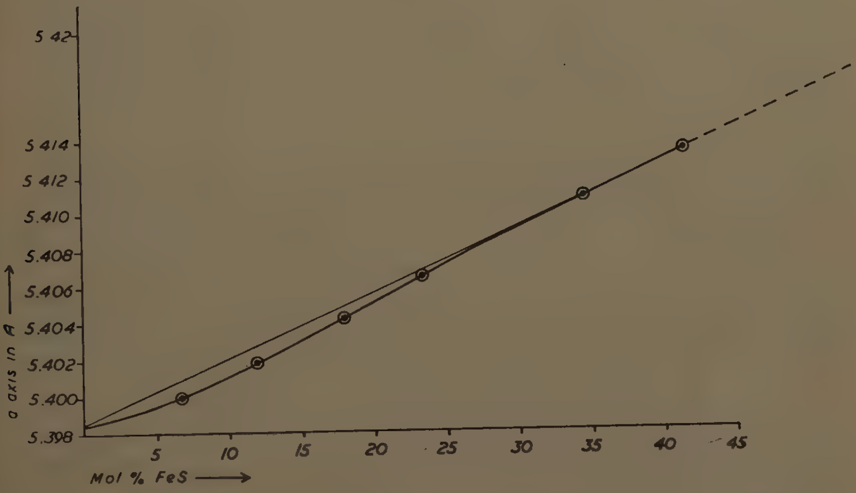

Fig. 5. Curve showing the relationship between the unit cell length and composition of β (Fe, Zn) S mix-crystals.

The a_o values so far known may be plotted against the FeS content of the mix-crystals. This has been done in figure 5.

The points plotted for mix-crystals containing less than 30 wt.% (34.47 mol.%) FeS are seen to lie below a straight line drawn between the a_o values for pure β ZnS and for mix-crystals containing 30 wt.% (34.47 mol.%) FeS. The biggest deviation occurs for mix-crystals with around 10 wt.% (11.98 mol.%)FeS. The distance becomes gradually smaller with increasing FeS content and mix-crystals with about 27 mol.% FeS are seen to lie on the straight line.

If one assumes the plot to continue as a straight line, one finds that the a_0 value of $5.413_{1\pm2}$A corresponds to 36 wt.% (40.84 mol.%) FeS. (As will be remembered from above this a_0 value was derived from runs at 890° C containing 40 wt.% (45.0 mol.%) FeS).

In order to check whether the above assumption is correct, two runs were conducted at 890° C. In both cases 36 wt.% (40.84 mol.%) FeS was mixed with ZnS. The runs lasted one and two weeks respectively. The unit cell length values calculated from x-ray diagrams were $a_0 = 5.4130$ and 5.4132 A. This proved that the a_0 versus composition plot continues as a straight line at least up to this point, and further that the maximum solubility of FeS in β ZnS at 890° C is 36 wt.% (40.84 mol.%).

The α-β Inversion Temperatures of the Mix-crystals.

A run at 900° C with a mixture consisting of 40 wt.% (45.0 mol.%) FeS and β ZnS produced α mix-crystals. From this and from the data obtained earlier, one may conclude that the maximum amount of FeS that can be dissolved in β ZnS is a little more than 36 wt.% (40.84 mol.%) and that the inversion point for mix-crystals of this composition is situated between 890 and 900° C.

A series of eight runs was then started in order to determine this inversion point. The first four runs were made with a mixture of 50 wt.% (55.1 mol.%) FeS and ZnS. This compound was first heated at 920° C (above the inversion point) for four days and later run for two weeks at temperatures and with results as shown in table III. The other four runs were made with 40 wt.% (45.0 mol.%)

TABLE III.

Duration of each run 2 weeks		Duration of each run 2 weeks	
Original α mix-crystals		Original β mix-crystals	
T° C of runs	Resulting modification	T° C of runs	Resulting modification
895	α	896	α
892	β	894	β
893	β	895	α
894	α	893	β

FeS mixed with ZnS. The compounds were first run at 880° C for three weeks so that homogeneous β mix-crystals were formed. Later these mix-crystals were used for runs also shown in table III.

For these runs special temperature controls were used so the uncertainty in the temperature measurements should not under any circumstances exceed ± 3° C. From table III the inversion point is seen to be 894 ± 3° C. In other words no α (Fe, Zn) S mix-crystal is stable below 894° C.

The β mix-crystals produced in four of the eight tabulated runs, were used for x-ray powder diagrams. The a_o values calculated from these were all inside $a_o = 5.413_{3\pm2}$A. Assuming the a_o versus composition curve in figure 5 to be a straight line, the obtained a_o value corresponds to a mix-crystal composition of 36.5 wt.% (41.36 mol.%) FeS. As a check on this assumption a mixture containing 36.5 wt.% FeS and ZnS was run at 894° C for seven days. Powder diagrams of the mix-crystal gave: $a_o = 5.413_{4\pm2}$A. (These data have already been presented in table 1). The a_o versus composition curve is thus a straight line from a composition of about 27 mol.% and through 41.36 mol.% FeS.

Because of the limited solubility it is not possible to find out the further course of this curve. However, if the straight line is extended towards full replacement of ZnS by FeS a ZnS type FeS with $a_o = 5.4345$A results.

The FG curve in figure 1 gives the temperatures at which β mix-crystals on heating will start to convert to the α modification, and at which α mix-crystals on "cooling" completely will convert to the β phase. In order to determine this curve, α mix-crystals of composition as shown in table IV were run from four to seven days at temperatures close to their inversion points.

The FD curve gives the temperatures at which α mix-crystals on "cooling" will start to convert to the β phase, and at which β mix-crystals on heating completely will change to the α modification. This curve was established by running β mix-crystals as shown in table V.

In table VI a compilation is presented of the inversion temperatures obtained from the twenty-five runs described in tables III through V.

The α-β inversion point for pure ZnS (upper column in table VI) was determined by ALLEN and CRENSHAW.[2] The limit of error in the

TABLE IV.
Runs with α mix-crystals.

Composition of α mix-crystals	T° C	Duration of runs	Resulting modification
10 wt.% (11.98 mol.%) FeS	995	4 days	α
	985	4 days	$\alpha + \beta$
	980	4 days	β
	982	4 days	$\alpha + \beta$
20 wt.% (23.48 mol.%) FeS	950	5 days	$\alpha + \beta$
	940	5 days	β
	943	5 days	β
	948	5 days	$\alpha + \beta$
30 wt.% (34.47 mol.%) FeS	920	7 days	$\alpha + \beta$
	910	7 days	β
	915	7 days	$\alpha + \beta$
	912	7 days	β

TABLE V.
Runs with β mix-crystals.

Composition of β mix-crystals	T° C	Duration of runs	Resulting modification
10 wt.% (11.98 mol.%) FeS	990	4 days	$\alpha + \beta$
	995	4 days	α
	993	4 days	α
20 wt.% (23.48 mol.%) FeS	965	5 days	α
	960	5 days	α
	955	5 days	$\alpha + \beta$
30 wt.% (34.47 mol.%) FeS	930	6 days	α
	927	6 days	α
	923	6 days	$\alpha + \beta$

other inversion point determinations in the same table does not exceed 4° C.

A modification may be present to a certain amount, depending upon its crystal symmetry, without showing up on the x-ray diagram.

TABLE VI.

Composition of mix-crystals	T° C for points on the FD curve	T° C for points on the FG curve
Pure ZnS	1020 ± 5	1020 ± 5
10 wt.% (11.98 mol.%) FeS	992 ± 4	980 ± 4
20 wt.% (23.48 mol.%) FeS	958 ± 4	945 ± 4
30 wt.% (34.47 mol.%) FeS	925 ± 4	913 ± 4
36.5 wt.% (41.36 mol.%) FeS.		894 ± 3

The phases in question are of a hexagonal and cubic symmetry respectively and thus certainly will be detected in the powder diagrams if they constitute as much as ten per cent of the total sample. It is possible that in some runs one phase has been present to the extent of less than ten per cent and has thus not been detected. Therefore the curve FG theoretically could be lying a little lower and the curve FD a little higher than indicated. The FGD field is rather narrow, meaning the inversion takes place completely over a narrow temperature interval. (maximum 13° C). Therefore, if in certain runs a phase was present to the amount of less than ten percent and thus not detected, the corresponding error in the inversion temperature of any mix-crystal would amount to 1.0 to 1.5° C at the most. This uncertainty is taken into consideration in the limit of error given in the inversion temperatures.

The Mechanism of the α-β Transformation.

In order to shed some light on the α-β transformation mechanism of the mix-crystals an extensive series of experiments were carried out.

The dimensions of the mix-crystals depend to a certain extent upon how long the compounds are heated. Mixtures with varying proportions of FeS and ZnS were heated for one week at temperatures slightly above the inversion points of the respective compounds. Thus α mix-crystals were produced, all of which seemed to be fairly equal in size, about 0.05 mm long. The crop of crystals produced in each run was then divided, by weighing, in two equal parts. While the one half was set aside for later runs, the other half was in each case sealed into a bomb and heated in vacuum for three more weeks at the same temperature as before the batch was split. The crystals grew in size during these three weeks and measured about 0.2 mm in length after completion of the runs.

The 0.2 mm mix-crystals from each run were then heated for four hours at temperatures of 50—60° C below their inversion points. In all cases it was then found that a complete transformation to modification β had taken place. The 0.05 mm mix-crystals (formed in the course of seven days) showed no signs of transformation after having been heated for four hours at temperatures of 50—60° C below their inversion points. Prolonged runs at the same temperatures showed in no case transformation after twelve more hours, but complete transformation in all cases after a total of twenty-six hours.

One here is comparing equal masses of mix-crystals, but the crystals in the one case are about: $\left(\frac{0.2}{0.05}\right)^3 = 64$ times as big in volume as in the other case. Consequently there are about 64 times as many "small" crystals to be transformed as there are "big" ones. In comparative runs with small and large numbers of mix-crystals of equal size it was found that a small number always converts much faster than a large number. DEHLINGER[29] showed that an Au-Cu single crystal only required a few minutes to transform from the regular to the tetragonal modification, in polycrystalline material the transition process lasted about ten hours. Explaining this phenomenon, DEHLINGER assumed that the transition in one crystal unit takes place immediately when at some point in the crystal a few atoms have arranged themselves in accord with the new symmetry. Under such an assumption the time necessary for the arrangement of the few atoms will be identical with the time of complete transformation. SHÔJI[20] showed on a theoretical basis how the transition in zinc sulfide could take place in the easiest way — the symmetry planes before and after transition being in a well defined position to each other. Since similar orientation relationships are observed in the metal transitions KRÖGER[30] expected that DEHLINGER'S ideas for the metals should apply also to pure zinc sulfide and showed it to be so.

While DEHLINGER and KRÖGER compared the time needed for transition of one crystal with that of polycrystalline material, one is in this case dealing with crystals that differ in number as well as in size. If all crystals were of the same size one should reasonably expect that the larger number would transform slower than the smaller number. Comparing equal masses, the total surface area of a large number of "small" crystals is many times bigger than that of a small number of "big" crystals. (The ratio being 16 : 1 in the earlier mentioned runs). An increased area of exposure should hasten the inversion at least if this is aided by a fluid phase in which the transforming material is soluble. The studied transformation, however, took place in a dry state, and is believed to proceed as a wave from centers that have already started to transform. Therefore, the fewer crystals present the more readily the transformation proceeds. (The ideal case being a single crystal transformation). In the runs with a great number of small crystals the transformation process was believed to be inhibited for the following two reasons: 1) A new center must arise in each grain of powder. 2) Heat must be transmitted through the intervening space from one crystal to another. This must take place chiefly by radiation from one crystal to its neighbors. Both these factors slow down the transformation. (BUERGER[31]).

Grinding of the wurtzite type mix-crystals was found to influence the intensities of the lines in x-ray powder diagrams. Thus after only two minutes of grinding the wurtzite lines were becoming diffuse, and getting even more so after four and eight minutes. The sphalerite [333] and [422] lines gradually appeared and were intense after thirty-two minutes. After sixty-four minutes of grinding, the pattern seemed to be entirely that of sphalerite. Grinding thus supplies the activation energy required to effect the transformation. The effect produced by grinding may also be produced by pressure and shear as shown by BRIDGMAN[32] who by such means irreversibly inverted wurtzite to sphalerite.

Further, a similar change in intensities as produced by grinding was observed on x-ray diagrams made after α mix-crystals had been heated 30—40° C below their inversion point for one, two, three, four, eight, and sixteen hours respectively.[1]

The ZnS α-β inversion involves no change in the first coordination of the atoms. The energy change must be accounted for by a change in the coordination of non-nearest neighbors. The transformation is sluggish as seen from the above described experiments. The α-β inversion for sphalerite has by BUERGER[31] been classified as a reconstructive transformation of a secondary coordination. The experiments carried out by the present author have verified BUERGER'S criteria for a transformation of this kind.

A Comparison of the Zn-S α-β Inversion with Liquid-solid Transitions.

Liquid-solid transitions have in recent years been investigated carefully, and some interesting results have been obtained. SCHAEFER[34] found for instance that tiny droplets of water (10—15 microns in diameter) might be super-cooled to a temperature of —39° C before inverting to ice. VONNEGUT[35] has studied the liquid-solid transition in tin particles. TURNBULL and CECH[36] studied the solidification of small metal droplets on a high temperature microscope stage. These investigations revealed that at least from a statistical standpoint the solidification temperature of smaller particles tends to be lower than that of larger ones. It has similarly been shown for solid transformations that large crystals invert at a higher temperature than small crystals, (LENNARD-JONES and DENT[37]). Undoubtedly not only the droplet size but also the time is of importance in deciding at which temperature solidification takes place. Thus if a small droplet solidifies instantly at a low temperature, it probably would solidify after some seconds, minutes or hours at a temperature somewhat closer to its melting point. In other words, if one has two chambers, the one containing comparatively big droplets and the other small droplets, and keep the droplets in both chambers super-cooled to the same degree, one would find that the larger droplets solidify first and then after some time the small ones. Thus there may be an analogy between liquid-solid inversions and α-β inversions like the one in ZnS.

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[1] Some of the x-ray powder diagrams made from these mix-crystals indicated a stacking disorder similar to that described by BARRETT[33] to occur in a number of pure metals. A disorder transformation has crystallographic characteristics substantially identical with the reconstructive type of transformation when the inversion is sluggish. It is therefore possible that the ZnS α-β inversion should be classified as a stacking disorder transformation instead of a reconstructive transformation.

TURNBULL[38] has derived a kinetic expression for the rate of transformation from liquid to solid by homogeneous nucleation as follows:

$$dV/dt = Iv\,(Vo—V).$$

Here V is the total volume transformed in the time t, Vo is the total volume at $t = 0$, I is the rate of nucleation per unit volume, and v is the volume of the droplet. The expression is deduced on the assumption that the nucleation has obtained a steady rate. It is important that the droplet size is uniform and accurately known.

The above equation may be integrated to give:

$$V = V_o\,(1—e^{—Ivt}).$$

As an experiment, one may now apply the last equation on the ZnS α-β transition. There were equivalent volumes (V_o) of 1/25 mm and 1/5 mm long particles. For the volume V_1 of small particles transformed in time t_1 one may write:

$$V_1 = V_o(1—e^{—I_1v_1t_1}).$$

For the volume V_2 of the larger particles transformed in time t_2, one may write:

$$V_2 = V_o(1—e^{—I_2v_2t_2})$$

When complete transformation of both small and larger particles has taken place, $V_1 = V_2$ and consequently:

$$1—e^{—I_1v_1t_1} = 1—e^{—I_2v_2t_2} \text{ or:}$$

$$I_1v_1t_1 = I_2v_2t_2$$

The rate of nucleation is proportional to the volume.[1] If I is the nucleation rate in the small particles, then the nucleation rate in the larger particles will be $I_2 = 64$ I. Remembering that $v_1 = (0.05)^3$ and $v_2 = (0.2)^3$ one obtains:

---

[1] This is an assumption made by TURNBULL for liquid-solid transitions. Nucleation of a new phase in a transformation in the solid state is a complex phenomenon, and as yet no general theory of it exists. The assumption made for liquid-solid transitions by the above mentioned author is here made for a solid-solid transformation as an experiment.

$$I \cdot (0.05)^3 \cdot t_1 = 64 \cdot I \cdot (0.2)^3 \cdot t_2 \text{ or:}$$
$$t_1 = 4^6 \cdot t_2$$

In other words, the time needed for complete inversion of the small crystals is according to this calculation, 4000 times longer than that required for the large crystals to transform to the β phase. Even if the result of the t_1/t_2 ratio calculation in no respect can claim to be more than approximate, it shows the same trend as the experimental data.

Lattice Dimensions of Pure α ZnS and of the α (Fe, Zn) S Mix-crystals.

From x-ray powder diagrams were calculated the unit cell a_o and c_o values for pure α ZnS as well as for the α (Fe, Zn) S mix-crystal series. (The rod thickness was as before $d = 0.30 \pm 0.05$ mm, and the same NaCl lines as earlier served as internal standard).

For α ZnS was calculated: $a_o = 3.814_{0\pm3}$A
$c_o = 6.257\,_{6\pm3}$A

The introduction of FeS into the α ZnS lattice causes an expansion in the lattice dimensions, very similar to that obtained for β ZnS. Table VII shows the relationship between unit cell dimensions and the composition of the mix-crystals.

TABLE VII.

Number of runs	FeS content of mix-crystals		T° C	a axis in A	c axis in A	Calculated molecular volume	Calculated specific gravity
	wt.%	mol.%					
3	0	0	1050	$3.814_{0\pm3}$	$6.257_{6\pm3}$	39.4160	4.1065
3	4.75	5.77	1020	$3.815_{6\pm4}$	$6.260_{7\pm4}$	39.4668	4.0821
3	9.30	11.18	990	$3.816_{6\pm4}$	$6.263_{8\pm4}$	39.5085	4.0595
5	22.00	25.72	980	$3.821_{9\pm4}$	$6.275_{3\pm4}$	39.6912	3.9902
3	30.00	34.47	960	$3.824_{7\pm4}$	$6.281_{6\pm4}$	39.7892	3.9485

The time required to obtain homogeneous mix-crystals in the runs tabulated above, appeared to be relatively short. A heating of

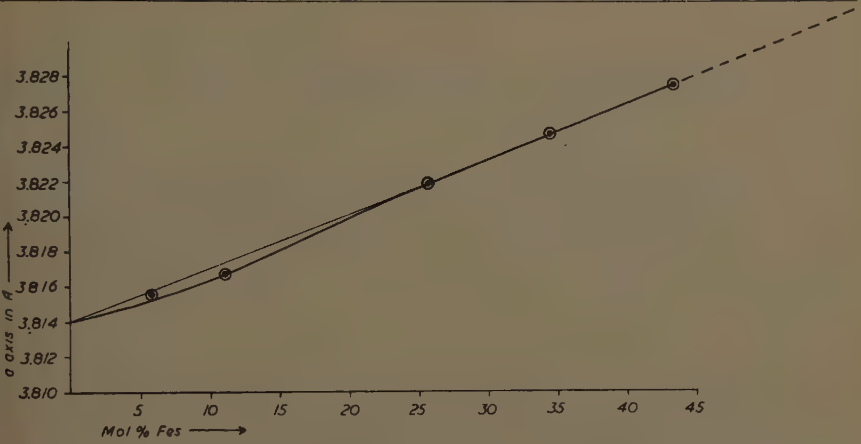

Fig. 6a. Curve showing the relationship between the unit cell a_o and composition of α (Fe, Zn) S mix-crystals.

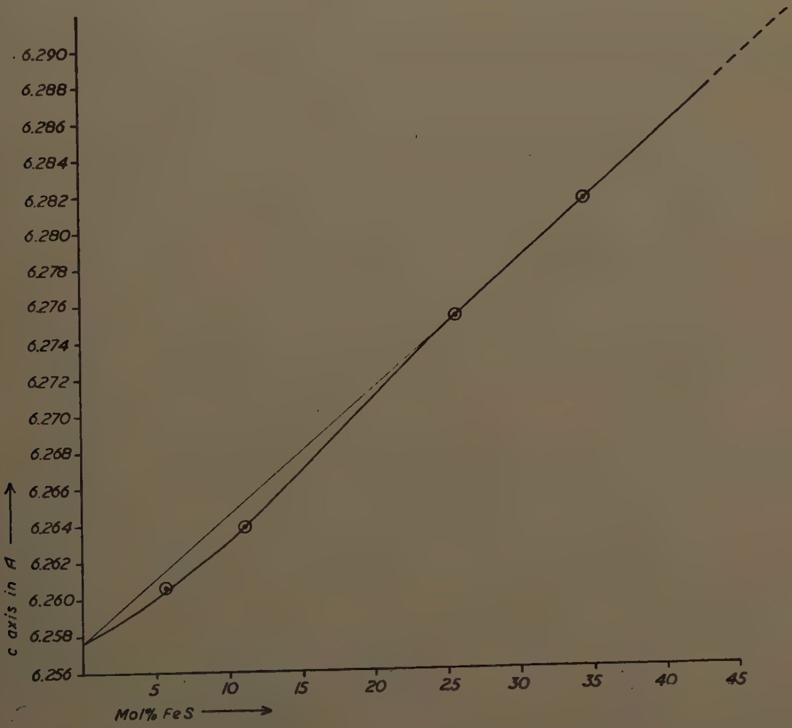

Fig. 6b. Curve showing the relationship between the unit cell c_o and composition of α(Fe, Zn) S mix-crystals.

forty-eight hours proved to be ample in all cases. (The same a_o and c_o values were obtained from runs lasting two days as from those lasting seven days). The a_o, c_o values respectively for the α (FeZn) S mix-crystal series are plotted versus mix-crystal composition in figures 6a and b. These plots show by comparison with figure 5 that the α ZnS behave analogously to the β ZnS when FeS is introduced into the lattice.

The Temperature-composition Relationship of the Equilibrium α Mix-crystals.

The CD curve (see figure 1) gives the equilibrium α (Fe, Zn) S mix-crystal composition.

In order to determine this curve two runs were conducted at 910° C. In the first of these runs 39.75 wt.% (44.74 mol.%) FeS; and in the scond 50 wt.% (55.1 mol.%) FeS were mixed with ZnS. The runs lasted seventy-two hours and the following unit cell lengths were calculated from x-ray powder diagrams: $a_o = 3.827_{7\pm4}$A and $c_o = 6.288_{0\pm4}$ for the first run, and $a_o = 3.827_{6\pm4}$ and $c_o = 6.288_{2\pm4}$ for the latter run. The curves shown in figure 6a and 6b are seen to be straight lines from a mix-crystal FeS content of 22 wt.% (25.72 mol.%) through 30 wt.% (34.47 mol.%). Assuming the curves to continue in this manner, the above given a_o and c_o values point to a mix-crystal composition of 38.5 wt.% (43.45 mol.%) FeS.

A mixture of this composition was run as a check at 910° C for seventy-five hours. The a_o and c_o values of the resulting α mix-crystals were well within the limit of error of the values given above. This shows that the curves in figure 6a and 6b continue as straight lines beyond the composition of 30 wt.% (34.47 mol.%) FeS.

In order to determine another point on the CD curve a mixture containing 50 wt.% (55.1 mol.%) FeS was run for forty-eight hours at 950° C. The unit cell dimensions of the mix-crystals thus formed were: $a_o = 3.828_{0\pm4}$A and $c_o = 6.288_{6\pm4}$. This corresponds to α mix-crystals containing 39.2 wt.% (44.17 mol.%) FeS.

If the solid solution were ideal, one now might, on the basis of the already known points on the CD curve, calculate the further course of the curve.[1] Such computations give for the FeS solubility about 46.0 mol.% at 1050° C and about 48.0 mol.% at 1170° C.

The Properties of the *α* and *β* Mix-crystals.

The physical properties of mix-crystals are functions of the properties of the end members of the mix-crystal series. Sometimes these functions are relatively simple. Thus GRIMM and HERZFELD[39] from the lattice energy studies by BORN and LANDÉ[40, 41, 42] derived a formula for the volume relationship existing between mix-crystals in a number of binary series.[2]

In figure 7 a are plotted unit cell length, molecular volume and specific gravity, each versus *β* mix-crystal composition. Figure 7b presents the same functions for *α* mix-crystals. The unit cell length versus mix-crystal composition curves are of such a nature that they cannot be expressed by a mathematical formula of the type derived by BORN and LANDÉ.

When mix-crystals of higher symmetry (as in the present case) are formed without deformation of the anions and without changes in the chemical bonds, the physical properties of the mix-crystals will vary nearly proportionally to the composition.

As will be seen later in this paper, the mix-crystal formation involves a change in the chemical bonds. This, therefore, explains why the BORN and LANDÉ formula is invalid in this case.

The curves presented in figures 7a and b are plotted from the data given in tables I and VII and extended by extrapolation towards complete replacement of ZnS by FeS.

In figure 7a one has on the left side *β* ZnS and on the right an FeS with the *β* ZnS structure. This hypothetical cubic FeS has the following characteristics: $a_o = 5.4345$ A, molecular volume about 40.125 A^3, and specific gravity $d = 3.63$.

Similarly in figure 7 b one finds on the left side *α* ZnS and on the

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[1] The background for these calculations will be discussed later in this paper.

[2] This work is well known and a further discussion of it here seems unnecessary. It may only be mentioned that both RETGER'S[43] and VEGARD'S[44] laws are special cases of GRIMM and HERZFELD'S more general formula.

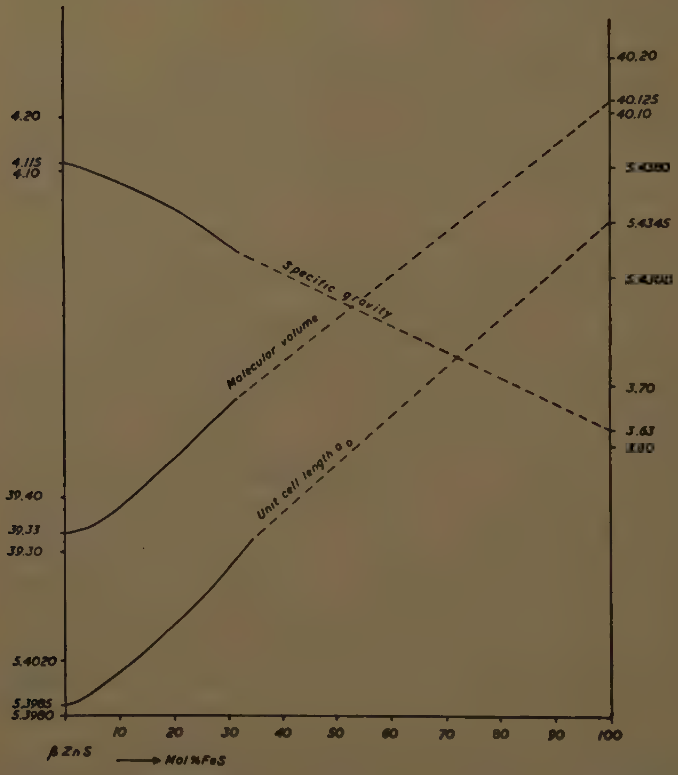

Fig. 7.a. Variations in the physical properties of β (Fe, Zn) S mix-crystals as functions of their composition.

right an FeS with the α ZnS structure. This hypothetical hexagonal FeS is earmarked as follows:

$a_o = 3.845$ A, $c_o = 6.328$ A, molecular volume about 40.46 A^3 and specific gravity $d = 3.60$.

The data given above are, without exception, obtained from rather daring extrapolations and must be regarded as approximations. Nevertheless, they are presented here because some later calculations are based upon them.

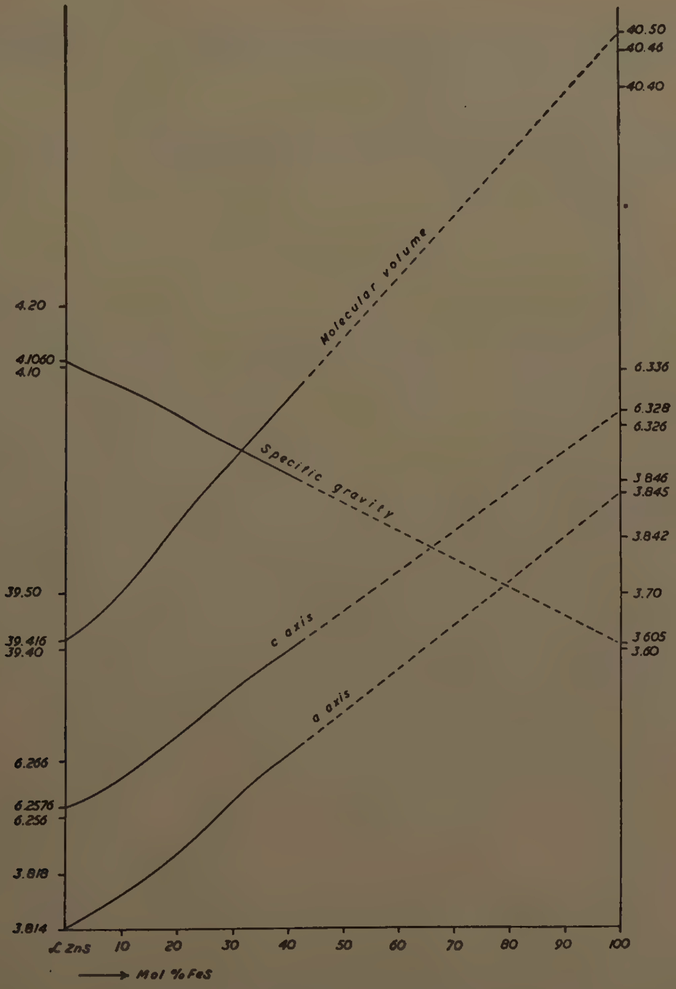

Fig. 7b. Variations in the physical properties of α (Fe, Zn) S mix-crystals as functions of their composition.

The Influence of Pressure upon the ZnS α-β Inversion.

The molecular volumes of β and α ZnS as well as of both types of mix-crystals have been calculated from x-ray data. (See tables I and VII, and figures 7 a and b).

The molecular volume of β ZnS is 39.333 A^3 and of α ZnS it is 39.416 A^3. The α-β transformation consequently is accompanied by a change in volume, $\Delta V = 0.083\ A^3$ for each molecule, or $0.083 \cdot 10^{-24} \cdot N = 0.05\ cm^3$ per mol. (Avogadro's number, $N = 6.02 \cdot 10^{23}$).

The Clapeyron-Clausius equation gives the temperature of transformation T, as a function of the pressure P:

$$dT/dP = T \cdot \Delta V/\Delta H,$$

where ΔH is the heat of transformation. The heat absorbed when β ZnS converts to the α modification is, $\Delta H = 3190$ cal/mol.[45]

Thus: $dT/dP = 1293 . 0.05/3190 = 0.0203$ degree/atm.

The dT/dP ratio has a positive value, since modification α has a greater molecular volume than modification β and since heat is absorbed.

Thus, under pressure, the β modification increases its stability field. At a pressure of one thousand atm., the inversion will take place around 1040° C. The molecular volumes for mix-crystals of any composition may be found in figures 7 a and b. The difference in the readings gives the wanted ΔV value. As an approximation, Δ H may be assumed constant for the mix-crystal series. Thus, when T as well as Δ V and Δ H are known, the influence of pressure upon the inversion point of mix-crystals of any composition may be calculated.

Magnetic Properties of the Mix-crystals.

While pure ZnS is diamagnetic, both the α and β mix-crystals are attracted by a magnet. The force of attraction is clearly a function of the mix-crystal iron content, and increasing with the latter. Unfortunately no quantitative measurements could be undertaken of the magnetic properties of the mix-crystals. It is believed, however, that if such measurements were done and the results plotted versus mix-crystal composition, curves very much like the ones presented in figures 5 and 6 would be produced. In other words there would exist a definite relationship between unit cell dimensions and the magnetic properties of the mix-crystals.

Reasons for the Increase in Unit Cell Dimensions due to the Introduction of Iron.

It has in the foregoing been shown that the introduction of iron into the zinc sulfide lattice causes an increase in the unit cell dimensions of both the high and low temperature modifications of ZnS. Such an increase might not be expected because iron both as ion and atom is smaller than zinc. WELLS[16] gives for the radius of Zn^{2+} in 4 coordination: 0.74A, and for Fe^{2+} in 6 coordination 0.75A. The approximate radius of Fe^{2+} in 4 coordination may be obtained from the last value by subtraction of about 5 %, which gives 0.71A. The radius of the Zn atom in 4 coordination is 1.31A (WELLS[16]) and the value given for Fe in 6 coordination is 1.23A. The atomic size of Fe in 4 coordination is somewhat smaller than the size in 6 coordination. How much smaller is not known, but one will assume its radius 1.20A. The corresponding values for sulfur in 4 coordination are 1.84A (ionic) and 1.04A (atomic).

From the earlier presented β ZnS lattice dimensions the Zn-S distance was calculated to be 2.337A.[1] When iron is introduced into the lattice the interatomic distance increases and by extrapolation of the curve in figure 5 one finds that when all zinc has been replaced, the distance, now between iron and sulfur is 2.353A.

From the values given above it is found by addition that the Fe-S distance, if the bonds are 100 % covalent, should be $1.20 + 1.04 = 2.24$A, and for 100 % ionic bonds $0.71 + 1.84 = 2.55$A. The Fe-S distance found by the extrapolation above is larger than the 100 % covalent distance and smaller than the 100 % ionic distance and suggests that the bonds have appreciable ionic character.

From thermochemical data, PAULING[18] and HAÏSSINSKY[46, 47] have calculated electronegativity coefficients for many elements. These values are such that the difference $X_A—X_B$ between the electronegativity coefficients for any pair of elements equals the square root of E_r, the ionic resonance energy of the bond A—B expressed in electron-volts. From the values of E_r, only differences $X_A—X_B$ are obtained, and it is necessary to assign an arbitrary value of X to one element. The choice of the absolute value of X was determined by the relation of electronegativity to ionization energy and electron affinity (WELLS[16]). Pauling plotted the known percentage ionic character for a number of compounds AB against the difference in electronegativity (X_A—X_B) and thus produced a fairly smooth curve showing this relationship.

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[1] The Zn-S distances in α ZnS were from the a and c values found to be 2.338 A and 2.346 A.

The electronegativities of the elements in question in this paper are for S : 2.5, Zn : 1.5. The X_A—X_B value for ZnS thus is: 1.0. Applying this value to Pauling's curve one finds that the ZnS bonds have about 20 % ionic character. This value cannot be taken literally, but will at least give some information about the ionic character of the mineral.

HAÏSSINSKY[46, 47] has pointed out that the electronegativity X of an element is not a constant under all circumstances. It depends on the valency and is not even constant for a given element in a particular valence state, but varies according to the compound from which it was derived. Thus the X value for Fe derived from pyrrhotite is likely to differ from the X value for Fe in FeS of the sphalerite or wurtzite structure type.

If one imagines a linear relationship between the increase in atomic distance (caused by iron replacing zinc) and the ionic character of the bonds in the (Fe, Zn) S mix crystals,[1] one finds that the bonds of the sphalerite or wurtzite type FeS must have approximately: $\frac{(2.353-2.24)}{2.55-2.24} \cdot 100 = 36\ \%$ ionic character. In other words, the expansion in the lattice parameters of ZnS may be explained, if one assumes that the percentage of ionic character of the bonds changes from about 20 % for pure ZnS to about 36 % for pure ZnS type FeS.

Changes in the bond types, as above described, have to the author's knowledge never before been applied to mix-crystals. It is believed that this method may be used successfully on many mix-crystal series for which the volume relationships cannot be mathematically expressed by means of a simple formula.

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[1] Since the length of the bonds are not proportional to their strength a linear relationship does not exist, but by making a thought experiment as above indicated, one at least is able to show that the bonds of the ZnS type FeS are of a more ionic character than the ZnS bonds.

Earlier in this paper a relationship between the magnetic properties and the unit cell dimensions of the mix-crystals was mentioned. The unit cell dimensions in turn are functions of the bond types. Consequently there exists a definite relationship between the latter and the magnetic properties of the mix-crystals.

The Temperature-composition Relationship of the Equilibrium β Mix-crystals.

Earlier in this paper it has been established that the maximum amount of FeS that can be dissolved in the β ZnS lattice is 36.5 wt.% (41.36 mol.%). Mix-crystals of such composition were produced at 894° C.

In order to determine a number of points on the GK equilibrium curve (see figure 1), an extensive series of runs was undertaken. Equilibrium mix-crystals were produced by solution as well as by exsolution. For the solution runs 40.0 wt.% (45.0 mol.%) FeS was mixed with ZnS, while for the exsolution runs mix-crystals containing 36.5 wt.% (41.36 mo.%) FeS were used.

The results of all these runs are compiled in table VIII.

The points between 800 and 890° C were obtained in the following way. Four bombs were run simultaneously at each temperature shown in table VIII. Two of these bombs contained a mechanical mixture of FeS and ZnS and the other two contained FeS rich α mix-crystals. (Both these compounds are described above). In the former two bombs mix-crystals were produced by solution, while in the latter two bombs mix-crystals of equilibrium composition were formed by exsolution.

After the four bombs had been heated for a period of time, the length of which depended upon the temperature, one bomb of each kind was taken out of the furnace. The mix-crystal composition was determined by means of x-ray diagrams. If there now in both bombs already were produced mix-crystals of identical composition, equilibrium had been established and the run could be interrupted.

Already the runs conducted at 880° C demonstrated that the exsolution process is more sluggish than the solution process.

While equilibrium at 800° C was reached in less than two weeks by solution and in about three weeks by exsolution, four weeks for

TABLE VIII.

The temperature-composition relationship of the equilibrium β mix-crystals.

T°C of solution and exsolution runs	a_o values in A from solution runs	required time for equilibr. by solution	Number of runs	a_o values in A from exsolution runs	required time for equilibr. by exsolution	Number of runs	FeS content of equilibr. mix-crystals	
							wt%	mol%
890	$5.413_{1\pm 2}$	48 hours	2	$5.413_{1\pm 2}$	48 hours	2	36.0	40.84
880	$5.412_{8\pm 2}$	48 hours	2	$5.412_{9\pm 2}$	72 hours	2	35.0	39.78
870	$5.412_{6\pm 2}$	72 hours	2	$5.412_{5\pm 2}$	120 hours	2	34.4	39.15
860	$5.412_{3\pm 2}$	105 hours	2	$5.412_{4\pm 2}$	7 days	2	33.7	38.40
850	$5.412_{0\pm 2}$	7 days	3	$5.412_{0\pm 2}$	10 days	2	33.0	37.67
800	$5.410_{9\pm 2}$	14 days	2	$5.411_{0\pm 2}$	21 days	2	29.8	34.25
750	$5.409_{4\pm 2}$	4 weeks	3	$5.409_{3\pm 2}$	6 weeks	3	26.7	30.89
700	$5.408_{4\pm 2}$	9 weeks	5	$5.408_{5\pm 2}$	12 weeks	5	24.0	27.93
600	$5.406_{3\pm 2}$	19 weeks	4	$5.406_{4\pm 2}$	28 weeks	4	19.0	22.35
500	$5.404_{3\pm 2}$	40 weeks	12	equilibrium not reached	more than 54 weeks	12	14.7	17.48

solution and six weeks for exsolution were required at 750° C. At 700° C the needed time was twice as long as that required at 750° C. This as well as the conditions at 600° C and 500° C are presented in table VIII.

For the runs conducted below 800° C a sufficient number of bombs of each type were prepared so one bomb of either kind could be taken out of the furnace and the material investigated each week, or at temperatures below 700° C each month. The variations in the a_o values thus were recorded from week to week (respectively from month to month).

At 500° C the solution process was complete after nine and one-half months. The a_o values obtained by means of x-ray powder diagrams after a total heating of eleven months, showed no increase beyond the values obtained after nine and one-half months. The exsolution runs at this temperature were extended to a total of twelve and one-half months. Certain interesting changes in the x-ray lines were recorded, but equilibrium mix-crystals were not obtained.[1]

In order to produce equilibrium β mix-crystals also below 500° C, methods different from those previously used had to be devised. Equilibrium was obtained at 400° C after three weeks by the following procedure. A compound consisting of 80 wt.% Zn and 20 wt.% Fe was heated at 850° C for twenty-four hours. This temperature was high enough to melt the compound. Upon quenching, the Zn-Fe alloy was crushed and ground to a fine powder. This powder was placed in a porcelain container inside a quartz tube, which again was stuck through a horizontal open-end furnace. The porcelain container was situated in the hot spot of the furnace, directly under the thermocouple. Through the quartz tube was then passed a mixture of 1 : 1 H_2S and H_2 gases.[2]

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[1] The material produced in the exsolution runs at 500° C showed, after having been heated for six months, separated double (333) and (511) lines instead of coinciding ones. The same happened to the (422) and (242) lines. If all normally occurring lines behaved in the same way one might have explained this by assuming that about equal amounts of mix-crystals of two different compositions were present. However, this was not so. The more probable solution is that one here is presented with mix-crystals of a deformed cubic structure.

[2] This ratio has been found by T. ROSENQVIST to be the right one for the formation of nearly stoichiometric FeS at 400° C.

After being heated for seven days at 400° C, x-ray powder diagrams showed that 50—60 % of the alloy had been converted to sulfide. The material was reground and put back into the furnace. After two more weeks of heating at the same temperature, all material was found to have transformed to sulfide. Calculations from x-ray powder diagrams of the unit cell length of the resulting β mix-crystals gave $a_o = 5.4021$ A. This corresponds closely to 11.0 % FeS by wt. (13.17 mol.%). Further heating of the substance produced no further increase in the a_o value.

Efforts to apply this method for the formation of β mix-crystals at 300° C were not successful. This temperature is too low for the sulfides to be formed within a reasonable length of time.

In order to obtain equilibrium mix-crystals in a comparatively short time at temperatures below 750° C a number of fluxes were mixed in with the sulfides. It was, however, experienced that the presence of any of these fluxes greatly influenced the equilibrium. A certain type of flux was found to entirely expel the FeS from the ZnS lattice.[1]

Thermodynamic Aspects of the FeS—ZnS System.

In the case of an essentially pure solute being in contact with its saturated solution, the equilibrium condition is that the chemical potential (partial molal free energy $\bar{F}$) for the solute is the same in the two phases. That is:

$$\Delta\bar{F} = \Delta\bar{H} - \bar{T}\Delta\bar{S} = O \ (1), \text{ or: } \Delta\bar{H} = \bar{T}\Delta\bar{S} \ (2)$$

where $\Delta\bar{F}$ is the difference in partial molal free energy of the pure and dissolved solute, $\Delta\bar{H}$ is the difference in partial molal heat contents of the pure and dissolved solute, and $\Delta\bar{S}$ is the difference in partial molal entropy of the pure and dissolved solute.

In the case of an ideal solution $\Delta\bar{S}$ is given by the equation:

$$\Delta\bar{S} = -R \cdot \ln(x) \ (3),$$

where x is the mole fraction of solute. If one introduces (3) into equation (2), one gets: $\Delta\overline{H} = T \cdot [-R \cdot \ln(x)]$ (4), or: $\ln(x) = -\Delta H/R \cdot 1/T$, or: $x = e^{-\Delta\overline{H}/RT}$ (5).

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[1] Further studies of the latter type of reactions are now progressing. The results may have significance for the dressing of iron-rich sphalerite ores and will appear in a later publication.

That means, if one plots the logarithm of x versus 1/T, one gets a curve, the slope of which is $-\Delta H/R$.

Thus in the FeS—ZnS system where FeS is the solute and ZnS the solvent, one may plot the logarithms of the experimentally found FeS concentrations versus 1/T. This has been done and the results are shown in figure 8. It will be noticed that the points for the con-

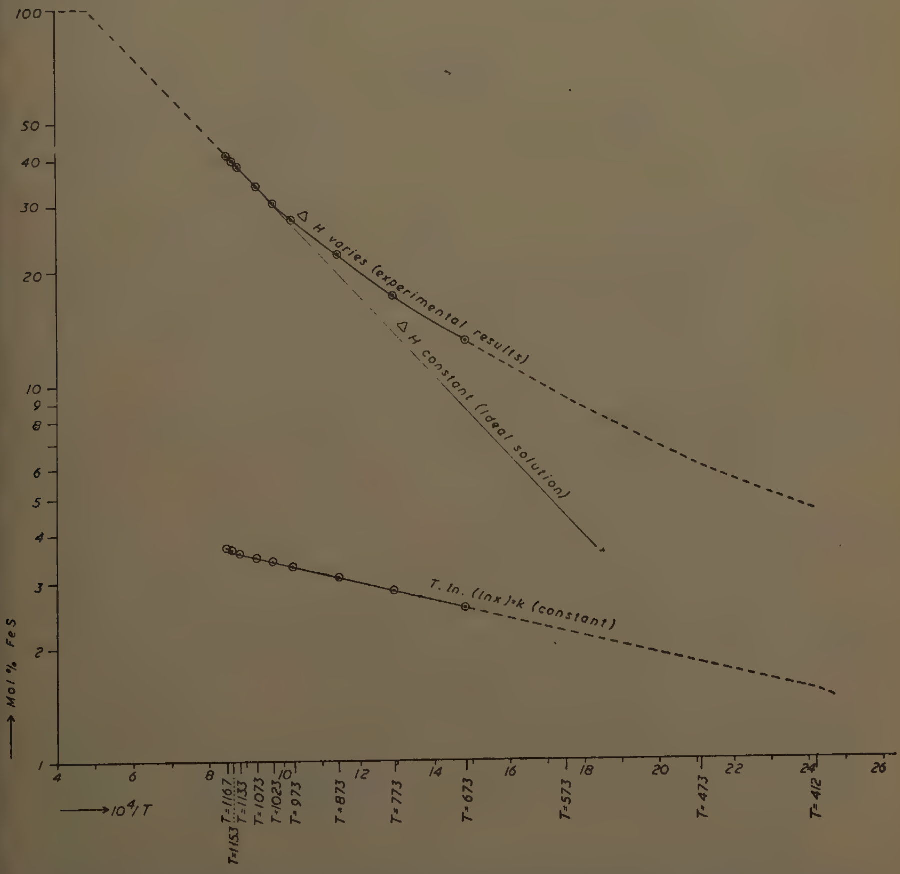

Fig. 8. Upper curve showing the relationship between 1/T and lnx. Lower curve showing the relationship between 1/T and ln(lnx). (In both cases x is the molconcentration of FeS in β mix-crystals.)

centrations between 41.3 mol.% FeS at 1167° K down to 30.89 mol.% at 1023° K lie very nearly on a straight line, in other words the slope of the curve is almost constant $= -\Delta H/R$, and the solution between these temperatures behaves like an ideal one.

Below 1023° K the points are seen to lie above the "ideal solution" values indicated by the straight line in the figure. The distances between this ideal curve and the plotted points increase markedly as the temperature decreases. In other words the solubility of FeS in ZnS is higher at temperatures below 1023° K than one would expect, if the solution were ideal. Ideal behavior at high temperatures and an increasing deviation from ideality at decreasing temperatures has been observed in liquid solutions. In such cases the deviation from ideality often may be expressed by simple equations which usually are of an exponential type. Thus, by plotting the logarithm of the logarithm of the FeS concentration against 1/T, it is not surprising that all the points may be joined by a straight line. (See lower part of figure 8).

Between 1167° K and 1023° K the ln(x) versus 1/T plot gives a straight line, the slope of which is $-\Delta H/R$. This slope may be expressed by the molconcentrations and temperatures thus:

$$\frac{-\Delta H}{R} = \frac{\ln(x_A) - \ln(x_B)}{1/T_A - 1/T_B} \quad (6) \quad \text{or:} \quad \Delta H = \frac{-R \cdot [\ln(x_A) - \ln(x_B)] \cdot T_A \cdot T_B}{T_B - T_A} \quad (7)$$

where A and B are two points on the straight line.

In (7) one may now insert the values: 41.36 mol.% FeS at at 1167° K, and 30.89 mol.% FeS at 1023° K. The result thus obtained is: $\Delta H = -4762.0$ cal/mol. By applying equation (2) one further finds: $\Delta S = -4.08$ cal/mol. degree.

Below point B the slope of the ln(x) versus 1/T curve diverges exponentially away from the constant $-\Delta H/R$ slope. To indicate that the $-\Delta H/R$ ratio is constant no more one may write: $-\Delta H'/R$ for the slope in any given point. In this expression $\Delta H'$ is a variable.

The relationship: $\ln(x) = -\Delta H'/R \,.\, 1/T$ (8) may now still be said to exist. By using the ln(x) and 1/T values for points situated close together on the GK curve, and assuming $\Delta H'$ to be constant between each pair of points, one may calculate the respective ΔH and ΔS

values. Table IX shows the variation in ΔH and in ΔS for different temperature intervals.[1]

TABLE IX.
Calculated heats and entropies of "mixing".

T°K	$-\Delta H$	$-\Delta S$
1167 — 1023	4762	4.08
1023 — 973	3983	3.89
973 — 873	3748.5	3.85
873 — 773	3297	3.78
773 — 673	2904	3.75
673 — 573	2615.5	3.88
573 — 473	2229	3.89
473 — 411	1737	3.67

It has been shown that the ln[ln(x)] values plotted versus the 1/T values produced a seemingly straight line. This may be written: $\ln[\ln(x)] = k \cdot 1/T$ (9), or: $\ln(x) = e^{k/T}$. If one inserts this in equation (8), the following formula may be be derived: $\Delta H' = e^{k/T} \cdot RT$ (10). The numerical value of k is —577.7.[2]

The linear relationship between ln[ln(x)] and 1/T is useful for the determination of points on the GK curve below 673° K (The lowest temperature at which equilibrium experimentally could be established). When the straight line in this plot is extended towards lower temperatures, one obtains for the solubility of FeS in ZnS the following values:

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[1] The formation of mix-crystals involves: 1) pyrrhotite changes its structure to that of ZnS, which implies an energy ΔH_1, 2) FeS with the ZnS structure enters the ZnS lattice. The energy required for the latter process is the heat of mixing ΔH_2. The ΔH mentioned above then is really composed of the energy required to transform FeS to the ZnS structure and the energy required for the latter type FeS to enter the ZnS lattice. Similarly the ΔS mentioned above is composed of the FeS transformation entropy and the entropy of mixing.

[2] By applying equation (9) on six known points on the curve in the lower part of figure 8, three independent determinations of k were done. The results varied from —577.62 to —577.76, average: — 577.7. In other words k is constant.

at 573° K 9.35 mol.% (7.75 wt.%) and at 473° K 6.17 mol.% (5.08 wt.%).

FeS converts at about 412° K its atomic arrangement from that of the Ni-As symmetry to that of the "supercell", the linear ln [ln (x)] versus 1/T relationship cannot be used for determinations of points on the equilibrium curve below this inversion point. The solubility at 412° K is from the plot in figure 8 found to be 4.66 mol.% FeS.

The energy released by this high-low inversion is 1250 cal/mol (Kelly[48]). The difference in energy between the high and low modifications of FeS will at the point L cause a break in the solubility curve as indicated in figure 1. How big the deviation in point L is between the GK and LK curves has not been determined.

Assuming the solubility curve to be ideal above the 894° C inversion point the lnx and 1/T values for the two already known points on the curve CD may be plotted against each other. The extension of the straight line through the two points determines the CD curve. The results of this plot have been mentioned earlier.

The FeS-ZnS diagram and all the results obtained from the numerous runs necessary for its establishment offer a great many possibilities for fascinating thermodynamic calculations. However, in order to keep the size of this paper within reason and not to get astray from the original purpose of providing a geologic thermometer, these will have to be considered at a later date.

Meteoric FeS.

FeS with a cubic symmetry has been described as occurring in iron meteorites. Fedorov[49] even went so far as to report this meteoric FeS to have the β ZnS atomic arrangement. Thus by replacing all Zn in β ZnS with Fe one obtains Fedorov's FeS. The physical properties of this FeS are found on the right hand side of figure 7a. Further, by extension of the curves in figure 8 towards full replacement of the Zn, one finds that the heat required would be about 9500 cal/mol at a temperature of around 2300° K. Comparing this temperature with the melting point of the Ni-As type FeS, and further by comparing the specific gravities of the same compounds, one may conclude that FeS of the β ZnS structure type does not exist in meteorites or on earth.

The Influence of Pressure upon the Solubility of FeS in β ZnS.

The mix-crystal composition in the FeS -- (Fe,Zn)S assemblage was shown above to be a function of the temperature. In these calculations, however, the influence of pressure on the FeS content of the mix-crystals was not considered.

Equilibrium mix-crystals containing x mol % FeS formed at temperature T and pressure P, will change in composition when the assemblage is exposed to a pressure P_1. The temperature remains unchanged. The composition x_1 corresponding to the new equilibrium conditions can be shown to be:

$$x_1 = x \cdot e^{\frac{\Delta F - \Delta F_1}{RT}} \qquad (11)$$

Here $\Delta F - \Delta F_1$ is the free energy change caused by the change in pressure from P to P_1. R is the gas constant. (0.082 liter atm./mol degree).

Now: $\Delta F - \Delta F_1 = \Delta V(P - P_1)$, where ΔV is the difference in mol-volume between pyrrhotite and ZnS type FeS.

The equation: $x_1 = x \cdot e^{\frac{\Delta V(P - P_1)}{RT}}$ (12) thus offers an opportunity for calculation of the influence of pressure upon the solubility of FeS in ZnS. It should be remembered that equation (12) has been deduced under the assumption that the solution is ideal. It has previously been stressed that the solution is not ideal. However, the error in the succeeding calculations of the influence of pressure upon the FeS solubility is not significant because the P-P_1 value is kept relatively small.

The molecular volume of β ZnS is 39.333 A^3 and that of FeS of the β ZnS structure is about 40.125 A^3 (figure 7a). The molecular volume of pyrrhotite is in the literature given from 29.6 to 29.9 A^3. Here will be used the mean value of 29.75 A^3.

Consequently $\Delta V = (40.125 - 29.75) = 10.375$ A^3, or:
$\Delta V = 62.46 \cdot 10^{-4}$ liter/mol.

In this ΔV is not taken into consideration the compressibility difference between hexagonal FeS and βZnS structure type FeS. The compressibility of most sulfides is insignificant at pressures of interest to this work. Further the compressibility of hexagonal FeS is likely of the same order of magnitude as that of the β ZnS type FeS. Therefore in this case the compressibility dependency of ΔV has only theoretical interest and will cause only minute errors in the calculations.

The ΔV value may now be inserted in equation (12) and by application of the T and x values of known points on the GK curve, the solubility-pressure relationships calculated. Table X shows the results of such calculations.

TABLE X.

Calculations of the influence of pressure upon the solubility of FeS in β ZnS by application of the equation: $x_1 = x \cdot e^{\frac{\Delta v(P - P_1)}{RT}}$ (12)

FeS solubility in mol.%

T°K	Laboratory conditions	1000 atm	2000 atm	3000 atm	4000 atm	5000 atm
412	4.66	3.88	3.22	2.69	2.22	1.84
473	6.17	5.25	4.48	3.79	3.22	2.75
573	9.35	8.17	7.17	6.26	5.47	4.80
673	13.17	11.70	10.49	9.35	8.35	7.45
773	17.48	15.80	14.30	12.95	11.74	10.64
873	22.35	20.49	18.73	17.15	15.73	14.40
973	27.93	25.80	23.82	22.07	20.36	18.84
1073	34.25	31.82	29.70	27.66	25.80	23.95
1167	41.36	38.66	36.23	33.90	31.70	29.76

The data given in table X are presented diagrammatically in figure 9. The influence of pressure upon the solubility is seen by no means to be neglible.[1]

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[1] The Tuttle apparatus at the Mineralogical-Geological Museum of the University of Oslo was used to check on the data given in table X for equilibrium mix-crystals formed at 500° C. Such mix-crystals were heated at 500° C under a pressure of 1000 atm for four weeks. The a_0 value of the resulting mix-crystals was smaller than that of the original mix-crystals and corresponded as closely as could be determined with the available x-ray powder cameras to about 16.5 mol.% FeS. This must be considered in fair agreement with the value of 15.8 mol.% FeS given in table X. Four weeks might not have been sufficient time for equilibrium mix-crystals to form. However, that unmixing takes place is established beyond doubt.

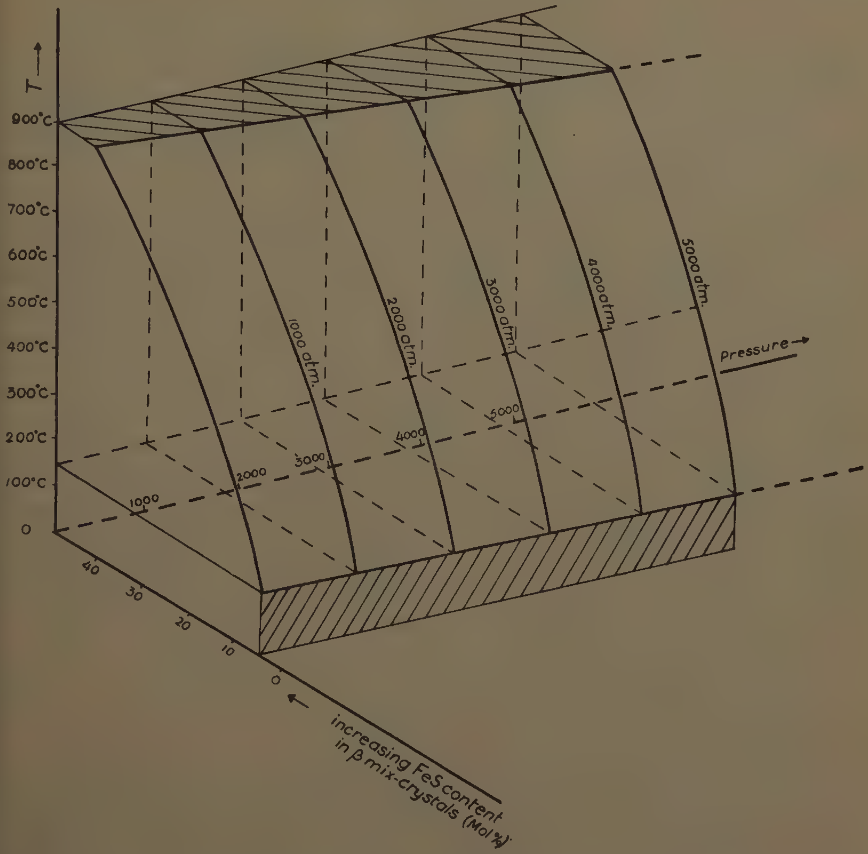

Fig. 9. Block diagram showing the relationship between temperature, pressure and composition of equilibrium β mix-crystals.

The FeS Activity.

The pyrrhotite which in nature forms mix-crystals with zincblende contains more sulfur (or rather, less iron) than indicated by the FeS formula. The FeS activity of pyrrhotite was expected to decrease with an increasing S : Fe ratio. Further information on this subject was obtained from T. ROSENQVIST,[1] who has investigated the FeS activity of iron sulfides over a wide temperature range.[2]

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[1] Institute for the Study of Metals, University of Chicago.

[2] The results of this research have not been published and the writer is grateful to ROSENQVIST for his kind permission to use them in this paper.

These detailed studies showed that the FeS activity of iron-sulfides of composition ranging from FeS to FeS_2 varies very little. The FeS activity of FeS_2 is less than 5 % smaller than that of stoichiometric FeS. However, when sulfur is added in excess of that indicated by the FeS_2 formula, a sharp drop in the FeS activity is recorded.

ROSENQVIST'S results agreed with those obtained from a number of runs with β ZnS and FeS and where extra S was added. These runs took place at 750, 800 and 850° C and lasted from 4 days to 3 weeks. When equilibrium was established the spacings were calculated from powder diagrams in the usual way.

In the cases where the excess sulfur was less than necessary to form FeS_2, the spacings obtained were nearly the same as those obtained in the previous runs where no extra sulfur had been added. However, when sulfur was added in excess of that necessary to form FeS_2 a sharp decrease in the FeS solubility was recorded. In the table XI the results obtained from a number of runs at 850° C have been compiled. As comparison is given the a_o value and FeS content of mix-crystals formed from ZnS and FeS without added sulfur.

TABLE XI.
Experiments carried out at 850° C.

Amount of excess S	No. of runs	Mix-crystal a_o value in A	Mix-crystal FeS content mol.%
None	5	$5.412_{0\pm2}$	37.7
Almost enough to form FeS_2	3	$5.411_{8\pm2}$	Average: 37.0
More than enough to form FeS_2	3	$5.407_{8\pm2}$	26.3

The $a_o = 5.411_{8\pm2}$ obtained from the runs containing almost sufficient sulfur to satisfy the FeS_2 formula indicates a decrease in the FeS solubility. (Compare with $a_o = 5.412_{0\pm2}$ obtained when no extra sulfur is added). Although the limit of error in the determinations of the a_o values $5.411_{8\pm2}$ and $5.412_{0\pm2}$ is so big that the values overlap, all the a_o values obtained from runs with excess sulfur were lower than those obtained from runs without extra sulfur. Assuming this difference significant the FeS solubility is seen to be lowered from 37.7 mol.%, when there is no extra sulfur present, to 37.0 mol.% when there is added sulfur to almost satisfy the FeS_2 formula.

This roughly corresponds to a derease of 2 % in the FeS activity and is seen to be in agreement with ROSENQVIST's results.

When more sulfur is added than necessary to form FeS_2, the solubility drops to 26.3 mol.%. Corresponding to the sharp drop in the FeS activity as found by ROSENQVIST. It may now be concluded that the FeS solubility in ZnS is not very sensitive to a variation in the S : Fe ratio in pyrrhotites and is only slightly (2 %) decreased when pyrite (FeS_2) forms mix-crystals with ZnS. Consequently the diagram (fig. 1) may be used as a geologic thermometer also where FeS_2 is found in equilibrium with (Fe, Zn) S mix-crystals.

However, if such a situation exists in nature, the diagram presented can not be used where sulfur is present in excess of the amount needed to satisfy the FeS_2 formula.

Exsolution of FeS from the (Fe, Zn) S Mix-crystals.

Earlier in this paper exsolution of FeS was shown to take place fairly rapidly at temperatures above 800° C. Below this temperature the exsolution process requires more time and becomes increasingly sluggish so that at 500° C equilibrium was not established even after 54 weeks. Solution at the same temperature needed 40 weeks to reach equilibrium conditions.

In the runs below 750° C sufficient FeS was unmixed from the mix-crystals formed at 890° C to show up on x-ray powder diagrams. The x-ray lines of this FeS did not coincide with the lines of normal NiAS type FeS. The crystal symmetry of unmixed FeS is much different from that of the NiAs type FeS.[1]

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[1] Further studies on this unmixing type of FeS are progressing. The unmixing mechanism was discussed orally with M. J. BUERGER who pointed out that this FeS must be expected to show a symmetry different from that of the NiAs type FeS. Determination of this structure may help explain the exsolution mechanism.

The data in the sixth vertical column of table VIII may be plotted on a logarithmic scale versus the temperatures given in the first vertical column of the same table. Thus a rough idea may be obtained of the time required for exsolution at temperatures below those at which equilibrium was experimentally established.

Figure 10 shows this plot. Complete exsolution is by extrapolation of the curve seen at 500° C to take place after 2 years. The corresponding lengths of time at 400, 300 and 200° C are around 5, 17 and 46 years. These values of course may not be taken literally because an extrapolation of the kind presented in figure 10 at best only will indicate the order of magnitude of the time periods needed for exsolution. However, even if the above guesswork values be multiplied a thousand times, exsolution will have had ample time to take place in many ore deposits. In large deposits formed at considerable depth the cooling process must have taken thousands of years. Studies of such ores have revealed that exsolution of FeS takes place to a very limited extent. The maximum amount reported does not exceed 2.0 %. The exsolved FeS occurs in lamellae parallel to the crystallographic directions of the sphalerite. STILLWELL[50] and later EDWARDS[51] studied such minute inclusions in sphalerite from

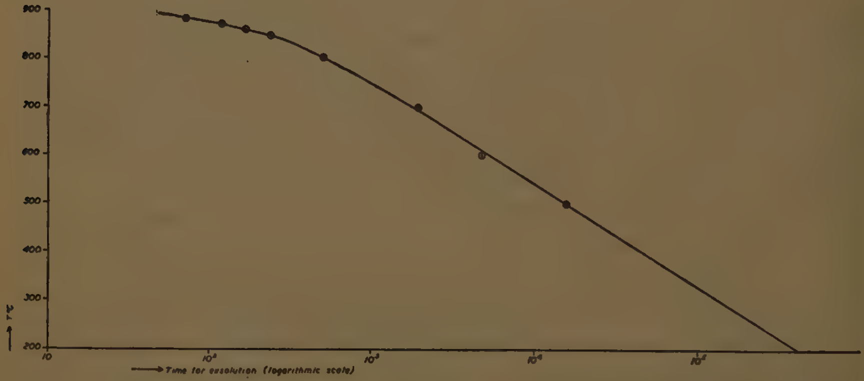

Fig. 10. Curve showing the time required for exsolution of FeS from β mix-crystals as function of temperature. The downward trend of the curve beyond the last point indicated is hypothetical.

Broken Hill. The amount of unmixed FeS in this ore was around 1.5 %. In most ores unmixing can not be detected at all.

As will be seen later, the Broken Hill ore was formed at temperatures well above 600° C. Even after the unmixed FeS is subtracted, the composition of the mix-crystals corresponds to a temperature above 600° C. The exsolution process in this ore started, but was arrested before the temperature had dropped to 600° C. This in spite of the fact that the cooling process of this ore must have lasted a great many years. Unmixing of FeS from (Fe, Zn) S mix-crystals in the laboratory was at 600° C completed in 28 weeks. The reason why unmixing takes place in the laboratory and not in nature, may be found when one takes into consideration the impurities frequently found in the mix-crystals in nature. Thus sphalerite whether it is associated with FeS or not, commonly contains manganese and/or cadmium. Both these elements may substitute for zinc in the same manner as iron does. It is possible that the presence of manganese and/or cadmium in the ZnS lattice may inhibit the exsolution of FeS.[1] The Broken Hill ore for instance contains besides iron, an appreciable amount of manganese.

Exsolution in alloys, some of which like the (Fe, Zn) S mix-crystals, occur in two enantiotropic modifications, may in certain cases be speeded up when homogeneous mix-crystals of the high temperature modification are used instead of mix-crystals with the low temperature atomic arrangement. The reason for this is that the inversion process with its rearrangement of atoms to suit the new symmetry, increases the mobility of the atoms and thus offers an opportunity for them to form mix-crystals of equilibrium composition. This principle was tried on the sulfide mix-crystals. α (Fe, Zn) S mix-crystals formed at 910° C and containing 38.5 wt.% (43.45

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[1] In an effort to shed some light on the influence of manganese on the FeS exsolution mechanism, a number of closed bombs containing finely ground Broken Hill ore are now being run at 550° C. The mix-crystal composition will be investigated monthly by means of x-ray powder diagrams. The results of these runs will be compared with the rate of unmixing taking place in other bombs being run at 550° C also, containing natural sphalerite rich in iron (as Broken Hill) but free of manganese and cadmium. Similar runs are later to be undertaken with sphalerite ores containing cadmium and no manganese.

mol.%) FeS were used. Five bombs were run at 500° C and five at 400° C. At 500° C the mix-crystals after several weeks inverted to the β phase, but equilibrium mix-crystals were not produced. At 400° the mix-crystals gave beautiful wurtzite x-ray diagrams even after having been heated for four months.

The Influence of Manganese and Cadmium on the Solubility of FeS in β ZnS.

In the foregoing, the FeS-ZnS equilibrium diagram has been presented. For ore deposits which contain only sulfides of FeS and ZnS, the diagram now may conveniently be used for temperature determinations. Most deposits however, contain manganese and cadmium dissolved with iron in the ZnS lattice. In most occurrences manganese is only a trace element, but in certain deposits (for instance Broken Hill) the manganese content may exceed 2 %. The cadmium content of most deposits normally lies between 0.1 and 0.3 %, but in certain cases exceeds 1 %. It was not known whether the presence of Mn and Cd had any measureable effect upon the FeS solubility and it was considered necessary to investigate their possible influence, before the FeS-ZnS equilibrium diagram could be generally used.

The (Mn, Zn) S Mix-crystals.

Manganese sulfide (MnS) occurs in three modifications; one pink MnS has the sphalerite structure, another pink MnS has the wurtzite structure, and a third MnS has the sodium chloride structure. The latter manganese sulfide is of a green color and is always the stable modification into which the two others will convert. The unit cell length of this green MnS is $a_o = 5.212$A (KRÖGER).[52]

When MnS is introduced into the ZnS lattice the sphalerite-wurtzite inversion temperature is lowered. The trend of this inversion curve was found from a number of runs with (Mn, Zn) S mix-crystals. The results of these runs are shown in figure 11. The composition of the mix-crystals as well as the modification obtained from each run are seen from the figure. It is interesting that α mix-crystals containing about 6.0 wt.% (7.6 mol.%) MnS are stable as far down as 765° C.

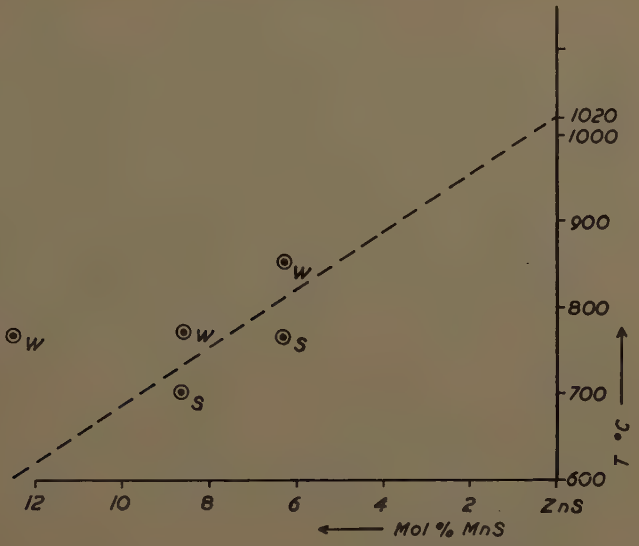

Fig. 11. Mean curve based on points indicated showing approximate relationship between α-β inversion temperature and composition of (Mn, Zn) S mix-crystals.

The (Mn, Zn) S mix-crystals, both of the sphalerite and wurtzite type are orange in color, the color being somewhat deeper for the mix-crystals richer in MnS. The solution of MnS in ZnS takes place much more rapidly than that of FeS in ZnS. Even at 765° C equilibrium (Mn, Zn) S mix-crystals were formed in 96 hours. KRÖGER[52] studied the solubility of MnS in the α ZnS and found that at 1180° C, 52 mol.% MnS may be dissolved in the α ZnS to form wurtzite type (Mn, Zn) S mix-crystals. KRÖGER[52] also showed that when mixing ZnS and MnS so the compound contains an excess of MnS, the spacing of this green MnS is not different from other green MnS. Therefore, if the green MnS dissolves ZnS, it can only be in an amount too small to be detected by his method.

In a number of runs green MnS was mixed with ZnS and heated to form β (Mn, Zn) S mix-crystals. The MnS dissolved in the ZnS caused a definite increase in the ZnS unit cell length a_o as seen from table XII below.

To insure complete solution each compound was run until no

TABLE XII.

MnS content of the mix-crystals		Unit cell length a_o in A.
wt %	mol %	
0	0	$5.398_{5\pm1}$
1.25	1.60	$5.399_{6\pm3}$
1.80	2.30	$5.400_{2\pm3}$
2.50	3.20	$5.401_{3\pm3}$
3.50	4.45	$5.403_{0\pm3}$
5.00	6.30	$5.405_{9\pm3}$

further increase in a_o occurred. In figure 12 the a_o values are plotted versus the mix-crystal composition.

The (Cd, Zn) S Mix-crystals.

Cadmium sulfide occurs in two modifications. One of these has the sphalerite structure and the other that of wurtzite. The

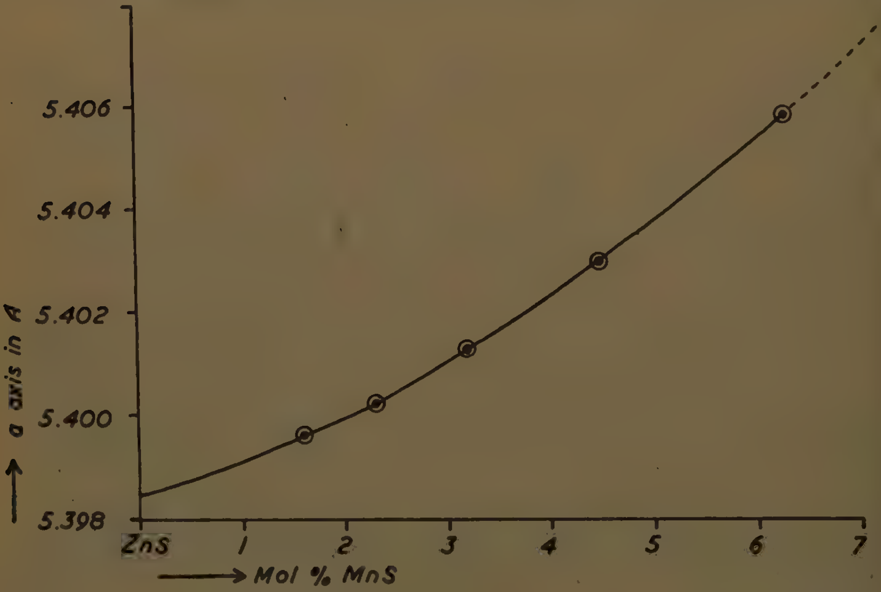

Fig. 12. Curve showing the relationship between the unit cell length and composition of β (Mn, Zn) S mix-crystals.

α ZnS and CdS are completely miscible (KRÖGER[53]). X-ray powder diagrams of the α (Cd, Zn) S solution series have been made by ROOKSBY.[54] The solubility of CdS in β ZnS, however, is limited. Mix-crystals containing CdS have a lower α-β inversion point than pure ZnS. In figure 13 is shown the approximate relationship between inversion temperature and mix-crystal composition. The general direction of the inversion curve was found from a number of runs with (Cd, Zn) S mix-crystals containing varying amounts of CdS. The runs were conducted at temperatures ranging from 750 to 1000° C.

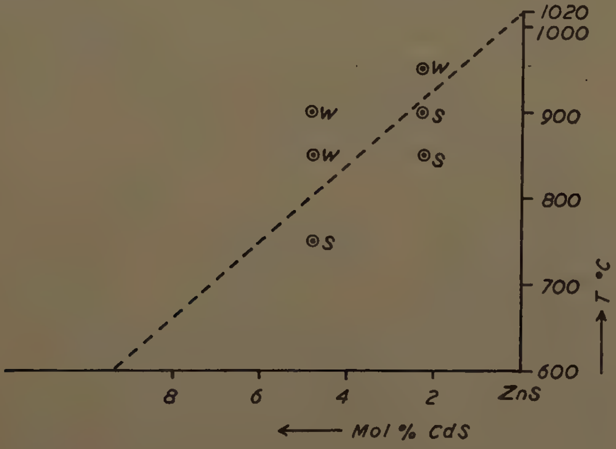

Fig. 13. Mean curve based on points indicated showing approximate relationship between α-β inversion temperature and composition of (Cd, Zn) S mix-crystals.

Mix-crystals containing 10 wt.% (4.8 mol.%) CdS are seen to invert at around 800° C.

Both the α and β type of the (Cd, Zn) S mix-crystals are brightly yellow in color.

Introduction of CdS into the ZnS lattice causes an increase in the unit cell dimensions. A series of runs with mechanical mixtures of ZnS and varying amounts of CdS gave the results shown in table XIII.

Complete solution was insured by the same procedure as applied for the (Mn, Zn) S mix-crystals.

TABLE XIII.

CdS content of the mix-crystals		Unit cell length a_o in A.
wt %	mol %	
0	0	$5.398_{5\pm1}$
1.10	0.50	$5.399_{4\pm3}$
1.50	0.70	$5.400_{0\pm3}$
2.50	1.15	$5.403_{0\pm3}$
5.00	2.30	$5.411_{2\pm3}$

The unit cell length versus composition relationship of the mix-crystals is shown in figure 14.

The (Cd, Zn) S mix-crystals, as was earlier experienced with the

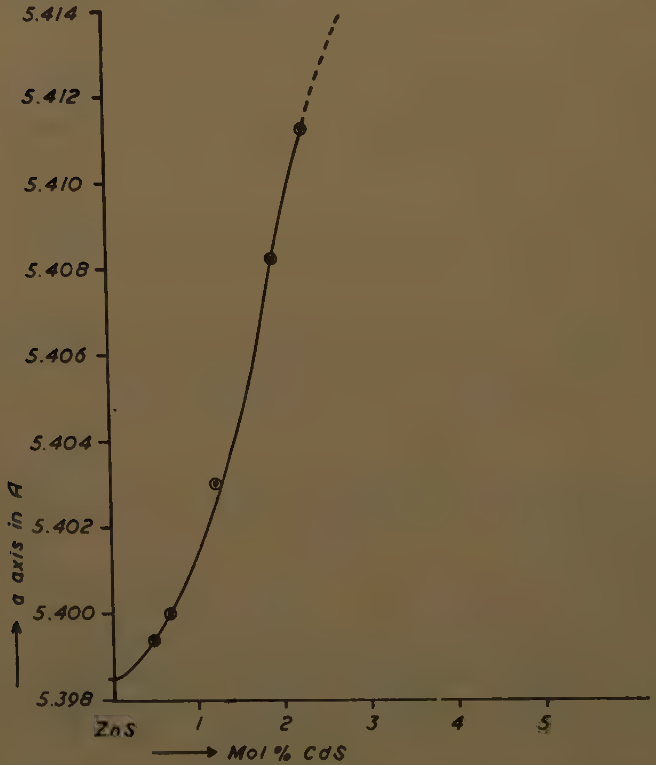

Fig. 14. Curve showing the relationship between the unit cell length and composition of β (Cd, Zn) S mix-crystals.

(Mn, Zn) S mix-crystals, formed at the same temperature in much shorter time than the (Fe, Zn) S mix-crystals.

FeS and MnS as well as CdS, when introduced into the ZnS lattice, all cause an increase in the unit cell dimensions. The increase caused by CdS is much larger than that caused by MnS. The latter increase in turn is much larger than that caused by FeS. Thus the increase obtained when 1.0 mol.% CdS is dissolved in the ZnS lattice is equal to that caused by 3.6 mol.% MnS or to 11.5 mol.% FeS. (See figures 5, 12 and 14). The α-β inversion temperature is similarly much more affected by CdS and MnS than by FeS. (See figures 1, 11 and 13).

The (Mn, Cd, Zn)S Mix-crystals.

Mix-crystals of modification β were formed from a mechanical mixture of CdS, MnS and ZnS. These compounds were heated at temperatures ranging from 750 to 850° C. The increase in the ZnS unit cell dimensions due to both CdS and MnS was calculated from x-ray powder diagrams and found, within the limit of error, to be identical to the sum of increases caused by the same amounts of CdS and MnS when these were independently heated with ZnS. For instance, a solid solution containing 1.8 wt.% (2.3 mol.%) MnS and 1.1 wt.% (0.5 mol.%) CdS in ZnS (heated for seventy-two hours at 800° C) had a unit cell length $a_0 = 5.401_{0\pm3}$ A. Previous runs (see tables XII and XIII) have shown that: 1) 1.8 wt.% (2.3 mol.%) MnS mixed with ZnS and heated below the α-β inversion point, produced mix-crystals with $a_0 = 5.400_{2\pm3}$ A. 2) Similarly formed (Cd, Zn) S mix-crystals containing 1.1 wt.% (0.5 mol.%) CdS had a unit cell length $a_0 = 5.399_{4\pm3}$ A. Thus the increase caused by MnS is: $5.400_{2\pm3} - 5.398_{5\pm1} = 0.001_{7\pm3}$ A and by CdS: $5.399_{4\pm3} - 5.398_{5\pm1} = 0.000_{9\pm3}$ A. Under the assumption of additivity the unit cell length of (Mn, Cd, Zn) S mix-crystals containing the above mentioned amounts of MnS and CdS will be: $a_0 = 5.398_{5\pm1} + 0.001_{7\pm3} + 0.000_{9\pm3} = 5.401_{1\pm3}$ A. This a_0 value is seen to be well within the limit of error of the $a_0 = 5.401_{0\pm3}$ value obtained from x-ray powder diagrams of (Mn, Cd, Zn) S mix-crystals of said composition.

The β (Mn, Cd, Zn) S mix-crystals are of a pinkish-yellow color.

The α (Mn, Cd, Zn) S mix-crystals have been studied by KRÖGER,[55] who presented the phase diagram for the ZnS-CdS-MnS wurtzite type mix-crystals. The α mix-crystals which contain much CdS and MnS are almost red in color. At 900° C all mixtures containing less than 48 mol.% manganese gave homogeneous products. When more manganese was mixed into the compounds, the runs resulted in red α (Mn, Cd, Zn) S mix-crystals and pure green MnS. Depending on the CdS content the MnS solubility varied from about 48.0 to 52.0 mol.% at 900° C. The presence of CdS therefore influences the MnS solubility very little.

The β (Mn, Cd, Fe, Zn) S Mix-crystals.

A compound consisting of 29.8 wt.% FeS, 1.8 wt.% MnS, 1.1 wt.% CdS and 67.3 wt.% ZnS was run at 800° C for 16 days. The β mix-crystals thus obtained were of a yellowish grey color. The unit cell length was calculated from x-ray powder diagrams and found to be $a_o = 5.413_{2 \pm 3}$ A. In table VIII was shown that equilibrium (Fe, Zn) S mix-crystal contain 29.8 wt.% FeS at 800° C. Such mix-crystals have a unit cell length $a_o = 5.410_{9 \pm 2}$ A. The increase in the cell length caused by the presence of 1.8 wt.% MnS is shown above to be $0.001_{7 \pm 3}$ A, and the increase caused by 1.1 wt.% CdS was $0.000_{9 \pm 3}$ A. By additivity we find that mix-crystals consisting of 29.8 wt.% FeS, 1.8 wt.% MnS, 1.1 wt.% CdS and 67.3 wt.% ZnS should have a spacing $a_o = 5.413_{5 \pm 3}$ A. The actual value is $5.413_{2 \pm 3}$ (see above). If now the presence of CdS and MnS have a tendency to decrease the FeS solubility, the x-ray a_o value should be lower than that found by additivity. Comparing the values it is seen that they are within the limit of error of each other. Therefore, if small amounts of CdS and MnS decrease the FeS solubility, the decrease is too small to be detected by the applied x-ray methods (and will be of the order of, at the most, 0.3 wt.% FeS).

Still the possibility existed that the presence of MnS and CdS might increase the solubility of FeS. In order to investigate this, a mechanical mixture consisting of 31.0 wt.% FeS, 1.8 wt.% MnS, 1.1 wt.% CdS and 66.1 wt.% ZnS was mixed and heated at 800° C for 16 days. The unit cell length of these mix-crystals calculated from x-ray powder diagrams was $5.413_{3 \pm 3}$ A.

If now the presence of MnS and CdS increased the FeS solubility to a measurable extent, an a_o value larger than $5.413_{3\pm3}$ A should result. Thus, if all the available FeS had gone into the solution a unit cell length of $a_o = 5.413_{8\pm3}$ A should have been expected. In a later run about 5 wt.% FeS more was added to the mix-crystals, and the compound was heated again for 15 days at 800° C. No further increase in spacing was observed.

From the above observations, it may be concluded that at least fairly small amounts of MnS and CdS will have no, or at least very little, influence on the solubility of FeS in β ZnS.

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PART II.

Introduction.

In the first part of this paper a geological thermometer based on the solubility of FeS in sphalerite was presented. Provided sufficient amounts of FeS are present to form equilibrium mix-crystals at the prevailing (P,T) conditions, the FeS content of the mix-crystals is a direct function of these conditions. In other words the FeS content of such mix-crystals gives the (P,T) function of formation.

Geological evidence concerning the pressure conditions existing during the mix-crystal deposition, may in natural occurrences often be obtained from other minerals formed simultaneously and under the same conditions as the (Fe, Zn) S mix-crystals.

The order of magnitude of the pressure thus known, the temperature of formation may readily be deducted. (It should here be remarked that an error of ± 1000 atm in the pressure estimate, leads to an uncertainty of about ± 25° C in the temperature determination).

In part two of this paper methods for determination of mix-crystal composition are described. Applications to geological problems are also considered. For this purpose numerous sphalerite hand specimens from various ore deposits were obtained through the kindness of numerous mining companies, museums, and private individuals. Only a few specimens were collected in the field by the author. Each sample was studied under the binocular lens, and the sphalerite (marmatite) carefully picked. The FeS content was determined chemically by Mr. H. B. Wiik, Helsingfors, Finland. The MnS and CdS contents as well as the amounts of a number of trace elements were determined spectrographically in cooperation with Mr. O. JOENSUU.[1]

X-ray powder diagrams were made of all samples.

Determinations of the Composition of the Ore Samples.

1. *Chemical Method*

The method used by Mr. WIIK for chemical determination of the iron content of the samples was in short as follows: The ground samples were fused by means of potassium pyrosulfate and dissolved in water and sulfuric acid. Thereafter, they were reduced with metallic zinc and sulfuric acid and finally titrated with potassium permanganate ($\frac{1}{20}$ N solution). With this method the average limit of error was $\pm$ 1.44 % of the iron content ($\pm$ 2.27 % FeS).

It was earlier mentioned that a limited amount of FeS sometimes unmixes from the mix-crystals during the cooling process. This unmixed FeS occurs in fine lamellae parallel to the crystallographic directions of the ZnS and too finely distributed to be seen through the binocular lens, much less be separated from the sample.

This FeS thus will be included in the chemical analysis. The temperature of formation deduced from the analytical result in such cases will be higher than that obtained from the x-ray unit cell length, which only tells how much FeS presently is dissolved in the ZnS lattice. Since at the most 2.0 mol.% FeS will unmix (and only from ores formed above 600° C), the difference between these two temperatures does not exceed 35° C. The highest value (the one obtained from chemical analysis) is the temperature of formation of the mix-crystals.

2. *Spectrographic Methods.*[2]

Much work was done in an effort to determine the content of FeS, MnS and CdS as well as of a number of trace elements in natural mix-crystals by means of spectrographic methods. The accuracy obtained by these methods was ± 5 % for the amount of FeS, ± 10 % for MnS and ± 5 % for the CdS content. The following description of the spectrographic methods is given by Mr. Joensuu.

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[1] The author wishes to stress that Mr. JOENSUU, Department of Geology, University of Chicago, being an expert on spectrographic analyses of this kind, did the major part of this work.

[2] Spectrographic analyses of trace elements in sphalerites have previously been performed by OFTEDAL[1], STOIBER[2], GABRIELSEN[3] and others.

Spectrograph: 21 foot grating instument by Jarrell Ash, Company. The dispersion in 1st. order 5 A/mm.
Power Source: Boliden Gruv A. B. Emitator, Sweden. 220 V. D. C. 7 amp. for the major constituents, 5 amp. for the trace analyses.
Electrodes: Anode: Central post. Undercutting type, turned out of 1/4 inch diameter Graphite rod (National Carbon Co. "Special', Grade) Cathode: 1/8- inch diameter carbon, pointed.
Arc gap: 4 mm.
Slit width: 0.05 mm.
Sector: Five steps with a step ratio of 1 : 2.
Exposure: For trace analyses: 30 seconds, for major constituents: 45 seconds.
Emulsion: Eastman Kodak 103—0 for the trace analyses, Spectrum analysis No. 1 for the major constituents.
Development: 3 minutes at 20° C in Eastman Kodak Developer D-19.
Photometer: Jaco non-recording by Jarrell Ash Company.

a. Determinations from Solids.

For the trace analyses one part of the sample was mixed with three parts of graphite; 20 mg of this mixture was packed in the electrode and arced.

Lines used:

Co	3453.505	5 — 1000 ppm
Ni	3414.765	5 — 1000 ppm
Ag	3382.891	1 — 100 ppm
Cd	3261.057	10 — 10000 ppm
Sn	3175.019	10 — 1000 ppm
Ga	2943.637	20 — 1000 ppm
Pb	2833.069	10 — 1000 ppm
Pb	2663.166	0.1 — 5 %
Ge	2651.178	1 — 1000 ppm

As internal standard was used:

Zn 2712.488

The plate was calibrated using a stepped iron arc spectrum. (PIERCE and NACHTRIEB[4]). For the calculations of the intensity ratios and for the background correction the calculating board was used. This has been described by HONERJÄGER—SOHM and KAISER.[5]

The standards were mixed as oxides with synthetic zinc sulfide. The working curves were linear. The precision of the method for germanium and cadmium is ± 5 %, for the other trace elements ± 10 % of the amount present.

The zinc sulfide content of the sample ranges from approximately 75 % to 100 %. The analyses were too high because zinc was used as internal standard. The results, therefore, were multiplied by the zinc sulfide percentage. The iron sulfide content was analysed and the zinc sulfide content could be calculated. (SCOTT[6]).

For the iron analyses one part of sphalerite was mixed with one part of cobalt sulfide and eight parts of graphite. Ten mg of this mixture was placed in a central post electrode and arced. The calculations were made as for the trace analyses but no background correction was made. For the construction of the working curves chemically analysed samples were used. The range of iron as sulfide (FeS) was 1—25 %. The precision of the method is about ± 5 % of the amount present. The lines used:

Fe 3018.982, Co 3013.596.

b. Determination by the Solution Method.

Twenty mg sphalerite was dissolved:

In a small vial 10 drops of the bromine-carbon tetrachloride mixture (2 : 1) were added to the sphalerite. After 5 minutes 2 drops of nitric acid (conc.) were added. After 5 minutes 4 drops of hydrochloric acid (conc.) were added. Bromine and carbon tetrachloride were evaporated at approximately 50° C. Then the acids were evaporated at approximately 90° C. The dry residue was dissolved in 2 cc of 3 % hydrochloric acid. This contained 0.2 mg chromium per ml. to serve as internal standard. For the analyses porous cups were used (FELDMAN[7]). The spectra were excited using a condenser discharge spark from the "Emitator".

Capacitance:	5 μF
Inductance:	100 μH
Resistance:	8 Ω

The presparking time was 10 seconds; the exposure time was 2 minutes.

The lines used:

Cu	3273.962	Cr	2751.871
Fe	2753.287	Cr	2751.871
Fe	2739.546	Cr	2751.871
Mn	2933.063	Cr	2751.871
Mn	2576.104	Cr	2751.871

The reproducibility was not good, however. The precision of the method was ± 10 %. The reason for this is probably the tendency of chromium, which was used as internal standard, to form a layer of chromium oxide on the electrode, thus hindering the liquid from getting through the bottom of the cup.

3. *X-ray Method.*

The x-ray technique has been described previously and it should not be necessary to repeat it all here. As will be remembered from this description, the amount of FeS in (Fe, Zn) S mix-crystals may, by use of x-ray powder diagrams, be determined well within the limit of error of ± 0.4 mol.% (corresponding to ± 0.0002 A in the unit cell length a_0). This uncertainty in the FeS determination again corresponds to a limit of error of approximately ± 7° C in the temperature determinations at temperatures above 400° C, and about ± 10° C at temperatures below 400°. The temperature determinations of mix-crystals containing only sulfides of iron and zinc, therefore, may well be done by application of the x-ray method. However, whenever manganese and/or cadmium also occur in the mix-crystals, the uncertainty attached to the determinations is increased. The limit of error of ± 0.0003 A in the unit cell length of (Mn, Cd, Fe, Zn) S mix-crystals, implies an uncertainty of about ± 0.6 mol.% in the FeS value. It is seen that when the FeS content of such mix-crystals is known, the x-ray method may serve well for determinations of the sum of the MnS + CdS content. On the other hand whenever the MnS and CdS contents are known, the FeS content may readily be determined from x-ray powder diagrams.

From the foregoing one may conclude that the most practical procedure for determination of the composition of mix-crystals of natural occurrence is:

1) Determination of FeS by the described chemical method.
2) CdS and MnS may be determined by any of the three methods. (The x-ray method can only be used after analysis on FeS has been performed).

   The most convenient way of finding the contents of CdS and MnS is by the spectrographic method.
3) The trace elements are determined spectrographically.

The Composition of the Ore Samples.

The ore samples were analyzed in the manner outlined above. The results are shown in table XIV.

The FeS contents are given in column 3 of table XIV. The amounts of MnS and CdS are further found in columns 4 and 5 respectively. Both MnS and CdS replace zinc in the ZnS lattice in the same manner as iron does. It has been found that small amounts of MnS and CdS will not influence the solubility of FeS in ZnS. Table XIV shows that the amounts of CdS and MnS in all the analyzed samples are too small to influence the mentioned solubility. Marmatites often are associated with chalcopyrite. The samples therefore also were analyzed on copper (Cu_2S) as seen from column 6. The copper thus detected was all assumed to occur in chalcopyrite. A correction of the FeS value corresponding to the amount of chalcopyrite was performed. Marmatites are often intimately connected with galena and in the cases where the marmatites are very dark (rich in FeS) it is hard to separate the two minerals quantitatively even with the help of a binocular lens. For this reason analyses on lead were also undertaken. All lead thus detected was assumed to occur in galena. The FeS value then was corrected accordingly. In column 15 are given the temperatures of formation of the mix-crystals obtained from the corrected FeS values. It should be noted that the temperatures listed in this column are not corrected for pressure. The pressure existing during the mix-crystal formation, must be obtained from geological observations. It is clear enough that this is a source of error. As shown earlier an error of ± 1000 atm. in

TABLE XIV.

1	2	3		4	5	6	7	8	9	10	11	12	13	14		15
Specimen no.	Locality	FeS		MnS wt%	CdS wt%	Cu_2S wt%	Pb ppm	Ag ppm	Ge ppm	Ga ppm	Ni ppm	Co ppm	Sn ppm	FeS corrected		T°C
		wt %	mol %											wt %	mol %	
40	*Broken Hill, Australia [8-17]	20.10	23.60	1.90	0.14	0.08	10	3	2	10	10	220	10	20.10	23.60	620
30	Mies, Bohemia[20]	4.04	4.91	x	0.8	x	280	100	33	10	x	x	95	4.04	4.91	150
43	Long Lake, Ont., Canada.	12.95	15.44	0.03	0.14	x	110	3	x	10	x	5	x	12.95	15.44	455
48	*Tetreault Mine, Quebec, Canada[21, 22]	10.82	12.96	0.10	0.19	x	7%	high	x	100	x	60	10	11.76	14.06	430
52	*Amulet Mine, Quebec, Canada[22,23,24]	13.50	16.08	0.12	0.13	0.5	450	100	9	55	x	160	70	12.95	15.44	455
53	*Calumet Island, Quebec, Canada[25,26,27]	12.40	14.80	x	0.19	x	5%	high	x	10	35	150	x	13.05	15.55	460
61	Renfrew Prospect, Ont., Canada	5.55	6.72	0.03	0.06	x	230	1	x	x	15	x	x	5.55	6.72	220
60	Buchans, Newfoundland [12, 22, 28] ..	5.35	6.48	0.01	0.21	0.40	10%	high	x	250	x	x	x	5.44	6.60	205
8	Alston Moor, Cumberland, England[29,30]	3.00	3.60	0.03	0.15	0.05	240	17	450	90	x	72	18	3.00	3.60	<138
15	Alston Moor, Cumberland, England [29,30]	2.60	3.17	x	0.18	x	150	80	530	x	x	48	x	2.60	3.17	<138
39	Foxdale, Isle of Man, England[31,32,33,34]	17.75	20.94	0.02	0.09	x	90	7	60	45	x	50	18	17.75	20.94	570
85	Germoe, Cornwall, Engl.[35]	4.60	5.59													180
88	—〃—	4.33	5.26													170
89	St. Agnes Mine, Cornwall, England[35,36]	9.92	11.91													370

*Localities which have received further attention in text.

1	2	3		4	5	6	7	8	9	10	11	12	13	14		15
Specimen no.	Locality	FeS		MnS wt%	CdS wt%	Cu_2S wt%	Pb ppm	Ag ppm	Ge ppm	Ga ppm	Ni ppm	Co ppm	Sn ppm	FeS corrected		T°C
		wt %	mol %											wt %	mol %	
4	Orijärvi, Finland[37]	16.60	19.63	0.50	0.16	0.08	3.8%	100	x	50	x	18	27	17.25	20.37	560
64	Kisko Ilijärvi, Finland[37] ..	14.10	16.77	x	0.14	0.25	800	15	x	45	x	100	x	13.83	16.46	480
5	Outokompu, Finland[38] ..	19.38	22.78	1.40	0.13	0.19	310	7	4	95	140	1000	100	19.17	22.54	600
10	Petsamo, Finland	3.85	4.69	x	0.54	x	230	8	32	190	x	48	x	3.85	4.69	145
12	Stanssvik, Finland	12.50	14.92	0.79	0.11	x	800	10	x	x	x	x	x	12.50	14.92	440
70	Petsamo, Peuravuono, Finland	8.85	10.65	0.28	0.38	x	170	15	230	130	10	120	x	8.85	10.65	335
72	Karjalohja, Kattelus, Finland.	14.75	17.52	0.32	0.14	x	10	1	x	10	10	70	x	14.75	17.52	505
73	Sortavala, Leppäselkä, Finland	1.90	2.32	0.02	0.10	x	500	x	x	x	10	100	x	1.90	2.32	<138
82	*Kimito Island, Finland [37, 39,40]	15.30	18.15	0.84	0.22	0.08	30					200		15.20	18.03	510
38	*St. Christoph, Breitenbrunn, (Erzgebirge), Germany[41,42]	18.65	21.96	0.30	0.38	x	10	9	x	10	x	50	25	18.65	21.96	595
42	—»—	18.28	21.54	0.60	0.28	x	x	3	x	x	x	55	10	18.28	21.54	585
69	—»—	7.17	8.66	0.07	0.41	0.16	14	x	50	15	15	450	x	7.00	8.45	280
65	Ivigtut Greenland[43]	17.55	20.71	0.04	0.18	0.22	1000	40	x	25	x	x	180	17.31	20.44	565
66	Marmiorilik, Greenland[43]	7.30	8.81	0.02		x								7.30	8.81	290
67	Tutup Agdlakofia, Greenland[43]	4.44	5.40	0.05	0.01	x	210	3	x	x	x	x	x	4.44	5.40	170
44	Morawizca, Banat, Hungary	5.86	7.09	x	0.27	x	10	22	2	25	x	x	x	5.86	7.09	235
18	Grua, Hadeland, Norw.[44]	20.40	23.93	0.36	0.24	0.14	170	6	2	21	22	260	45	20.25	23.76	625
19	—.—	20.10	23.60	0.40	0.25	0.08	x	5	2	15	18	260	18	20.00	23.48	620

TABLE XIV.

1	2	3		4	5	6	7	8	9	10	11	12	13	14		15
Specimen no.	Locality	FeS		MnS wt%	CdS wt%	Cu_2S wt%	Pb ppm	Ag ppm	Ge ppm	Ga ppm	Ni ppm	Co ppm	Sn ppm	FeS corrected		T°C
		wt %	mol %											wt %	mol %	
21	Melkedalen, Ofoten, Norway[45]	13.10	15.72	0.28	0.13	0.07	10	2	19	160	x	13	x	13.02	15.52	455
23	Sextus, Røros, Norway[46,47]	18.90	22.24	0.04	0.19	x	32	20	x	17	40	70	10	18.90	22.24	595
24	Storvarts, Røros, Norway[46,47]	12.80	15.28	0.09	0.14	0.05	700	33	x	30	18	26	10	12.75	15.21	450
25	Mofjell, Rana Norway[48]	7.50	9.05	0.06	0.20	x	3.1%	15	x	90	16	x	x	7.74	9.34	300
57	—»—	9.40	11.30	0.07	0.17	x	10	x	x	x	x	x	x	9.40	11.30	350
28	Svenningdalen, Helgeland, Norway[54]	9.70	11.65	x	1.04	0.05	3.2%	100	x	40	10	x	x	9.97	11.97	370
27	Jakobsbakken, Sulitjelma, Norway[49]															480[1]
86	Bjørkåsen, Ofoten Norway,[50]	12.32	14.71													440
29	*Kafveltorp, Sweden[51]	16.30	19.29	0.06	0.11	x	5.0%	100	x	90	10	27	90	17.15	20.26	560
31	*Falun, Sweden[51]	20.18	23.68	0.40	0.15	x	5.0%	100	x	80	10	30	160	21.24	24.87	645
58	*Saxberget, Sweden[51]	5.60	6.78	0.02	0.15	x	6.5%	high	x	x	10	70	50	6.05	7.32	245
63	Krigstjärn, Sweden	7.00	8.45	0.41	0.15	x	450	8	x	x	x	x	x	7.00	8.45	275
1	Berg Prospect, Wrangel, Alaska, U.S.A.	16.95	20.03	0.13	0.54	x	470	25	19	x	17	18	46	16.95	20.03	555
3	Sedanka Island, Alaska U.S.A.[52]	3.39	4.13	0.01	0.20	x	220	7	x	x	x	200	x	3.39	4.13	<138
7	*Balmat, N.Y., U.S.A.[53]	8.30	10.00	0.02		x										320
59	*Edwards Zinc Mine, N.Y., U.S.A.[53]	8.90	10.70	0.07	0.06	x	300	10	6	140	x	x	x	8.90	10.70	335
87	—»—	8.80	10.60													330
36	Roxbury, Conn., U.S.A.	7.85	9.47	0.01	0.10	x	250	8	73	140	10	130	60	7.85	9.47	305

1	2	3		4	5	6	7	8	9	10	11	12	13	14		15
Specimen no.	Locality	FeS		MnS wt%	CdS wt%	Cu_2S wt%	Pb ppm	Ag ppm	Ge ppm	Ga ppm	Ni ppm	Co ppm	Sn ppm	FeS corrected		T_oC
		wt %	mol %											wt %	mol %	
41	Gilman, (Red cliff) Colorado, U.S.A.[55]	14.70	17.46	0.09	0.23	x	1.8%	75	x	120	x	x	20	15.00	17.80	505
45	Cape Rosler, Mo., U.S.A.	4.95	6.01	0.05	0.18	x	450	8	2	170	x	x	x	4.95	6.01	190
46	Warren, New Hampshire, U.S.A.	8.00	9.64	0.35	0.10	x	5.0%	100	x	150	x	x	x	8.49	10.23	325
54	Austinville, Virginia, U.S.A.[56,57,58]	2.45	2.99	0.01	0.18	x	480	7	30	20	x	x	x	2.45	2.99	<138
81	Gray Mine, Illinois, U.S.A.[61,62]	2.77	3.38	0.01	0.18	x	580	40	50	40	10	x	x	2.77	3.38	<138
75	Hoskins Mine, Wisconsin, U.S.A.[59,60,62]	6.22	7.52	0.01	0.05	x	450	4	110	x	x	x	x	6.22	7.52	245
76	Tiffany Mine, Wisconsin, U.S.A.[61,62]	4.70	5.71	0.01	0.08	x	200	14	160	10	x	x	x	4.70	5.71	180
77	Thompson Mine, Wisconsin, U.S.A.[61,62]	0.91	1.11	0.01	0.14	x	480	16	40	60	x	33	x	0.91	1.11	<138
78	Rodham Mine, Wisconsin, U.S.A.[61,62]	1.62	1.98	0.01	0.10	x	1100	100	x	20	x	54	x	1.62	1.98	<138
79	Shullsbury district, Wisconsin, U.S.A.[59,60,61,62]	3.68	4.48	0.01	0.13	x	100	15	70	35	x	20	x	3.68	4.48	~135
80	—»—	5.23	6.34	0.01	0.10	x	450	70	100	30	10	80	x	5.23	6.34	205
56	—»—	4.20	5.10	0.01	0.10	x	5.0%	15	150	x	x	x	x	4.45	5.41	170
83	Joplin, Missouri, U.S.A.[60,61,62]	1.54	1.88	x	0.45	x	35	x	120	14	x	x	x	1.54	1.88	<138

the pressure estimate leads to an uncertainty of ± 25° C in the final temperature reading. If another system, in addition to the one already completed, was worked out, and mix-crystals of both types were formed under the same (P, T) conditions, one could easily find the exact pressure and temperature of formation.

After having determined the composition of the two types of mix-crystals, one could make a pressure versus temperature plot for each of the two mix-crystal types. (See figure 16 for plot of the Broken Hill ore). Where the curves intersect, one then read off the (P,T) values of formation.

It would of course, be of great value to know both the pressure and temperature existing during the formation of such mix-crystals, and the author hopes in the future to be able to work out another system for this purpose.

Some of the samples did not contain both iron sulfide and sphalerite. Other samples which carried both minerals did not clearly demonstrate that the iron sulfide was present in sufficient amounts during the mix-crystal formation. In all such cases, where equilibrium conditions might not have existed, the values given in column 15 of table XIV must be considered minimum temperatures.

The sequence of deposition, or paragenesis, of the ore minerals in a particular deposit can be determined in part from hand specimens, but a full understanding of the processes involved, can be obtained only from a study of the mineral textures as revealed in polished sections under the microscope.

A detailed description of all the deposits from which the sixty-three samples listed in table XIV originate, would have to be compiled almost entirely from already published accounts and not from the results of the author's own field investigations.

For this reason, and also in order not to increase the volume of this publication unduly, only a few deposits will be briefly described as examples of the applicability of the FeS — ZnS thermometer.

Applications of the FeS—ZnS Thermometer to Geological Problems.

The FeS — ZnS geological thermometer may be used to shed light on a number of problems of geological significance. Many examples of its applicability for determinations of the temperature of formation of ore bodies are found in table XIV.

Ore bodies are often surrounded by zones of skarn minerals and among the latter marmatite sometimes occurs. In such cases one may determine the temperature of formation of the ore body as well as of the skarn minerals. An example of this is found in table XIV (specimens 38, 42, 69). In many ore bodies more than one generation of marmatite occurs. The samples 25 and 57 as well as 23 and 24 indicate two generations of mix-crystals. Thus assisted by detailed field work and by application of the FeS — ZnS system one may get a good picture of the different stages of metamorphism which ore bodies have undergone.

The temperature of formation of pegmatites has long been a topic of great interest to geologists. In table XIV are found results from determinations of the temperature of formation of mix-crystals in pegmatites.

Geologists have paid great attention to the problems of granitization. In the succeeding paper in this journal KULLERUD and NEUMANN have applied the FeS — ZnS system for determination of the temperature of granitization in an area in Northern Norway.

An important factor in the classification of ore deposits, is the temperature of their formation. Based upon geological information obtained through field work and often coupled with a certain amount of guesswork, an ore body is said to have been deposited within one or other wide temperature range, given names by LINDGREN and others. The FeS — ZnS system offers an opportunity for pinning down the actual temperatures of formation of the ore deposits and thus also sharply defining in degrees the temperature ranges expressed by terms like pyrometasomatic, hypothermal, mesothermal, etc.

These are some of the applications of the FeS — ZnS thermometer. The possibilities for applying the system to other problems are many and fascinating. It is hoped that the thermometer will be a useful tool to geologists in determining geological temperatures.

Examples of Temperature Determinations.

1. Ore Bodies.

a. Broken Hill, New South Wales, Australia. This area is well known to most geologists all over the world, because of the numerous geological publications issued in connection with its ore deposits. The temperature of formation of the ore bodies has been discussed a number of times among geologists but so far no absolute temperature value has been agreed upon.

It is, therefore, natural that this problem again is discussed in the present paper. The following geological description of the area is compiled from earlier publications.

The ore deposits are confined to the oldest rocks of the region. These rocks, the Willyama series, consist of a shield of ancient sedimentary rocks, basic lavas and intrusive igneous rocks which have suffered erosion. The shield is considered, provisionally, to be Lower pre-Cambrian in age. It is a grand geological complex consisting principally of intensely folded and metamorphosed clays, shales, sandstones and allied types, together with a notable development of acid and basic igneous material (ANDREWS[8]).

A major mountain-building period, associated with igneous intrusions, succeeds the deposition of the sedimentary series. The rocks were folded and intruded by numerous large dikes of dolerite. The peculiar nature and disposition of the intense metamorphism to which the Willyama series has been subjected, suggests that the Broken Hill Shield occupies the site of an area of intense metamorphism produced by great local heating at a considerable depth from the surface and conditions of great pressure. This localized area of intense alteration may be called the Broken Hill Hearth. The main "lode" occurs within the longer axis of this zone of most marked metamorphism. Its trend conforms, definitely, to the local strike of the associated sedimentary and igneous rock exposures, namely, north-east and south-west approximately. These associated rock assemblages comprise a magnificent suite of augen gneiss, platy gneiss, basic lavas (amphibolites), sillimanite-garnet-mica-feldspar gneisses, massive garnetized and granitized sediments. The lodes are massive lead-zinc sulfide replacement ore bodies forming (before erosion) a long continous, irregularly shaped, flat, curving pencil of ore roughly 2000—3000 feet high and 300 feet thick. In a longitudinal section, the deposit describes a broad arch flat in its middle (highest) portion, and pitching downward at each end. One of the most significant features of the ore is that its boundaries are almost invariably parallel to bedding planes in the ancient metamorphosed sedimentary strata enclosing it.

There is considerable range of opinion regarding the classification of the ore deposit according to its temperature and depth of formation (GUSTAFSON[9]). ANDREWS[10, 11] took the extreme view that the lodes as well as the igneous rocks of the district were injected under conditions of great temperature, great pressure, and deep burial as "pressure lenses" in "pressure slacks". LINDGREN[12] classed the deposits as hypothermal. GUSTAFSON[13] believes that folding occurred at great depths characteristic of plastic deformation, perhaps as deep as 20—30 km, whereas ore deposition occurred long after folding, following an interval of uplift and erosion at much shallower depths, and further that the lodes owe their shapes to replacement of folded rocks. The first conspicuous act of ore deposition was the introduction of large amounts of "lode pegmatite" characterized by green microcline and quartz, accompanied by silicification and garnetization of the gneiss. Much, if not all, of the quartz and feldspar of this stage replaced the rock rather than intruded in igneous fashion. In this early stage, the introduction (or recrystallization) of large quantities of rhodonite, johanssenite, pyroxmangite, bustamite, hedenbergite, and garnet, all relatively "high temperature" minerals, occurred. Considerable amounts of pyrrhotite, arsenopyrite and loellingite occur in the ore, often closely associated with hedenbergite. These mineral associations denote relatively high temperature of ore deposition. The large size of the deposit, the general absence of zoning, and the limited extent of wallrock alteration are characteristic of hypothermal deposits as defined by LINDGREN[12] and GRATON.[14]

The temperature of formation of the ore deposit has been discussed by MAWSON,[15] who found blue lode quartz from Broken Hill to be the "low" type and believed that ore deposition occurred below 575° C. SCHWARTZ[16] considers that ore deposits containing cubanite-chalcopyrite intergrowths are formed above 400—450° C. STILLWELL[17] described such intergrowths from this deposit. Both these "geological thermometers" mentioned above, are pressure conditioned.

The ore listed first in table XIV is from Broken Hill. In this sample, dark marmatite, some chalcopyrite, and a little pyrrhotite are visible to the naked eye. The sample is believed to be representative, and since free FeS is found with the marmatite, there is reason

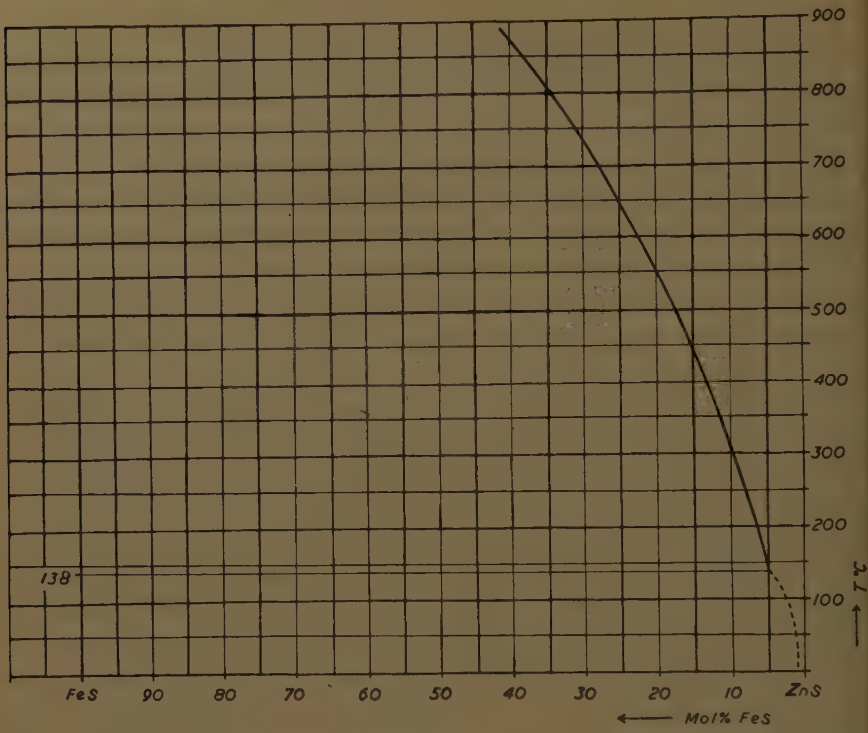

Fig. 15. Graph showing the composition of equilibrium β (Fe, Zn) S mix-crystals as a function of temperature.

to believe that the marmatite contains as much FeS as corresponds to the (P,T) conditions existing during the mix-crystal formation.

Figure 15 shows the part of the FeS — ZnS system (figure 1) of interest for the determinations of the temperatures of formation of natural mix-crystals.

The Broken Hill specimen contains 20.1 wt.% (23.5 mol.%) FeS. This composition corresponds to a temperature of 620° C. This of course is provided the ore formation took place at atmospheric pressure. Figure 9 gives information of the (P,T) relationship for mix-crystals containing any given amount of FeS.

The upper curve in figure 16 shows the pressure versus temperature plot for (Fe, Zn) S mix-crystals containing 23.5 mol.% FeS. The temperature of formation of the Broken Hill marmatite is found somewhere on this curve.

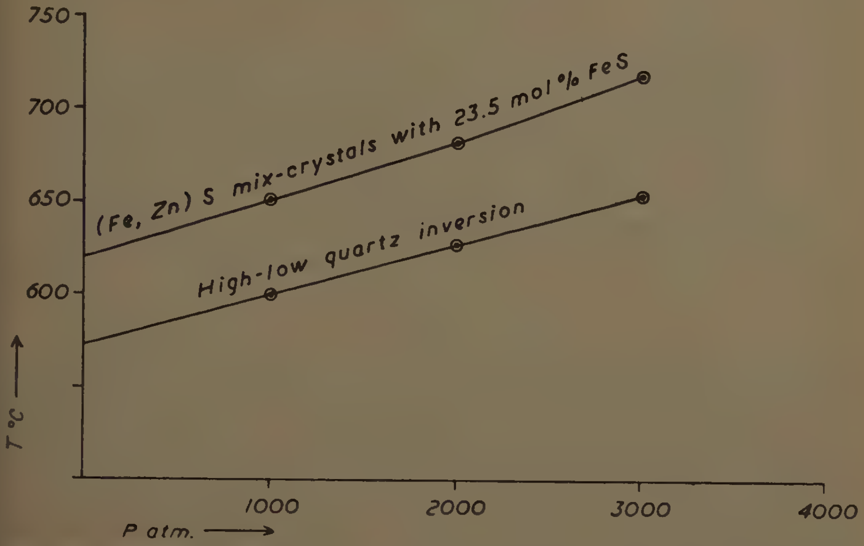

Fig. 16. Comparison between the temperature-pressure relationship for (Fe, Zn) S mix-crystals from Broken Hill (upper curve) and the temperature-pressure relationship for high-low quartz inversion (lower curve).

Geological evidence discussed by GUSTAFSON[9] indicates that the ore in question may have formed at a depth of about 8 km corresponding to a pressure of about 2500 atm. On this assumption, the marmatite of Broken Hill is seen from figure 16 to have been formed at 700° C.

The lower curve in figure 16 shows the variation in the high-low inversion temperature of quartz as experimentally found by GIBSON[18] and YODER[19]. The inversion takes place at about 645° C at a pressure of 2500 atm. As was earlier pointed out, MAWSON[12] found the lode quartz to be of the "low" type. This indicates that the quartz has been formed at a lower temperature than the marmatite.

b. *Amulet Mine, Quebec, Canada.* In the Noranda area, western Quebec, a number of ore deposits are being worked, one of which is Amulet. Here rhyolite, rhyolite flow breccias, and tuffs are intruded by sills and dikes of quartz diorite, all of pre-Cambrian age. These rocks are closely folded and sheared and are intruded by granite and by great diabase dikes.[23, 24]. The ore carries pyrite, chalcopyrite, pyrrhotite, sphalerite, etc. with very little gangue. (EMMONS[22]).

The specimen from this locality contains sphalerite, pyrrhotite, and chalcopyrite. The analysis (see table XIV) indicates a temperature of 455° C before pressure correction. It is hard to make a definite statement concerning the pressure at the time of mix-crystal formation. However, assuming the pressure to have been 2000 ± 1000 atm., the temperature of formation is found (from figure 9) to be 505 ± 25° C.

c *Balmat — Edwards, New York, U. S. A.* At Edwards, and at Balmat,[53] N.Y. a garnet gneiss including limestone bodies is found below a thick limestone series above which is a later garnet gneiss. Intruding these are gabbro, amphibolite, syenite, and granite, all probably pre-Cambrian. Great deposits of zinc ore are opened both at Edwards and Balmat. They replace limestone and generally follow the bedding or banding of this. The primary ore minerals are pyrite, sphalerite, galena, chalcopyrite, pyrrhotite and barite. Primary sphalerite is of a dark chocolate-brown color. BROWN[53] analyzed primary sphalerite both from Edwards and Balmat Mines. The results of these analyses agree very well with those presented in table XIV of samples 7, 59, and 87.

Assuming that during the ore formation there existed a pressure of 2000 ± 1000 atm., one finds that the temperature of formation of the Balmat marmatite was 375 ± 25° C and for the Edwards marmatites 385 ± 25° C.

2. Skarn.

St. Christoph, Erzgebirge, Germany. The ore deposits in Erzgebirge, Germany have been described in detail by REH[41] and KOCH.[42]

Devonian schists and limestones were intruded by large granitic masses. The main sulfide ores are found on the border between the granite and the country rock. The intrusion of granites resulted in development of skarn minerals. Zinc sulfide is a common constituent of the garnet type of skarn.

Samples 38 and 42 (see table XIV) are taken from two of the main ore bodies in the area, while sample 69 is from the zone of garnet skarn.

The pressure existing during the formation of the ores and the skarn has been appreciable, maybe as high as 2500 atm. If one corrects the temperature readings for a pressure of 2500 ± 1000 atm., one obtains the following temperatures for the formation:

1) Main ore bodies: Sample 38: 655 ± 25° C.
Sample 42: 645 ± 25° C.

2) Skarn: Sample 69: 345 ± 25° C.

3. Pegmatites.

a. *Tetreault Mine, Quebec, Canada.* The Tetreault Mine,[21] about 52 miles west of Quebec, lies in an area of quartzites and gneiss intruded by amphibolite and granite pegmatite. A lens of impure dolomitic limestone in gneiss dips 65 degrees east. The ore-bearing member lies near the pegmatite. The limestone is replaced by tremolite rock and by ore composed of sphalerite, galena, pyrite, calcite and micas. (EMMONS[22]).

The ore sample from Tetreault Mine is composed of a dark marmatite intimately mixed with galena; a little pyrrhotite and specks of chalcopyrite are also visible to the eye. The analysis of this ore (table XIV) shows that 7.0 wt.% Pb was present in the sample. Assuming all Pb to occur in galena and correcting the analysis accordingly, one finds that the sample contains 11.76 wt% (14.06 mol.%) FeS. Not corrected for pressure, this FeS content corresponds to a temperature of 430° C.

Pegmatites are believed to have formed generally at depths of 1.5 km or more (EMMONS[2]). This depth corresponds to a pressure of approximately 500 atm. If one assumes a pressure of 1000 ± 500 atm., one finds from figure 9 that the marmatite was deposited at a temperature of 455 ± 15° C.

b. *Calumet Island, Quebec, Canada.* The zinc-lead deposit is located on Calumet Island, Quebec, 53 miles northwest of Ottawa. The complex ores resemble those of Edwards, New York, and Mountau ban, Quebec, and other Grenville deposits. They occur in moderately inclined Grenville metasediments, including crystalline limestone, amphibolite, amphibolite-gneisses, quartzite (rustyweathering), gneisses, and various granitized gneisses (migmatites).

The sulfide minerals comprise pyrite, sphalerite, pyrrhotite, galena, tetrahedrite, chalcopyrite, and marcasite, deposited in the order named. The sulfides are leanly disseminated in all the rock types in the ore zones, and lenticular masses largely deposited in carbonate and contact metamorphic rocks. They replace carbonate, quartz, and all silicates except serpentine. The association of the ore minerals with quartz and contact silicates indicates that they accompanied the same agents that produced silicatization and silicification. The association of sulfides with late stage processes in pegmatite like material is evidence that they formed a phase of the pegmatitic or volatile-rich granite magma. This "pegmatitic" material is considered to have been responsible for the contact metamorphism and migmatiziation as well. (MOORHOUSE,[25] ARMSTRONG,[26] FITZ.[27]) Assuming a pressure of 1000 ± 500 atm. to have existed during the formation of this ore, the pressure-corrected temperature of formation is 485 ± 15° C.

c. *Kimito Island, Finland.* The pegmatites from Kimito have been described by ESKOLA[37] and PEHRMAN.[39] KNORRING[40] analyzed a sample containing a manganese garnet enclosed in an iron-rich sphalerite from this locality. The result of Knorring's analysis of the sphalerite is found in table XIV. The present author has received a specimen of this sphalerite. Iron-sulfide is not present in this sample. However, PEHRMAN[2] stresses the occurrence of both pyrrhotite and pyrite with the sphalerite. The temperature of 510° C given in table XIV for this sample, should be corrected for pressure. Assuming a pressure of 1000 ± 500 atm. at the time of formation, one arrives to a temperature of 535 ± 15° C.

4. Sphalerite of More Than One Generation.

Kafveltorp, Falun, Saxberget, Sweden. The ores in these localities are of the so-called Falun type.

The deposits of this type have originated in connection with the metasomatic alteration which has been called "magnesia metasomatism" and is intimately connected with the intrusion of the oldest granites (urgranites) of Central Sweden. These granites intruded the folded leptite formation (metamorphosed volcanics and sediments). The ore bodies occur mainly in skarn-dolomite zones, or in quartzite. They contain pyrite, chalcopyrite, pyrrhotite, sphalerite etc. (MAGNUSSON[51]).

The specimen from Kafveltorp, gave as seen from table XIV, a temperature reading of 560° C. The pressure existing during the ore formation is hard to estimate, but hardly exceeded 2000 atm. Assuming a pressure of 2000 ± 1000 atm., the temperature of formation of this ore is found to be 620 ± 25° C.

The specimen from Falun gave a temperature reading of 645° C. Assuming the same pressure as above, one arrives at a temperature of 700 ± 25° C for the formation of this ore.

In Saxberget mine there are two distinctly different types of sulfide deposits. The one type is similar to the ores described from Kafveltorp and Falun, and the other occurs in a straight fissure filled with a fine-grained, compact mass of sphalerite, pyrrhotite, pyrite, chalcopyrite, and galena. The specimen from Saxberget was collected from the latter type of ore. This specimen is, in table XIV, found labelled with a temperature of 245° C. Correction for the influence of pressure (2000 ± 1000 atm.) brings the temperature of formation of this ore up to 295 ± 25° C.

Discussion of the Results of the Temperature Determinations.

On studying table XIV, one finds that the FeS content of the analyzed samples varies considerably. The largest amount of FeS (almost 24 mol.%) occurred in the specimens from Grua, Hadeland, Norway and from Broken Hill, Australia. The smallest amount of FeS (about 1.0 mol.%) was found in the specimen from Thompson mine, Wisconsin, U.S.A. The temperature field, within these extremes, ranges from 625° C down to some temperature well below 138° C. The temperatures recorded are not corrected for the influence of pressure. Even if one corrects all temperatures for pressure, the answers obtained will not in all cases give the temperature of formation of the mix-crystals.

Undoubtedly a number of the investigated ores have been formed under non-equilibrium conditions. As previously pointed out, field work followed by microscopic studies of the individual specimens can alone give definite information about the paragenesis of the ore minerals.

Non-equilibrium conditions seem to have existed during the mix-crystal formation in the Joplin deposit. The amount of FeS dissolved in the ZnS varied in three specimens from less than 1.0 to about 1.9 mol.%. All specimens were composed of a light colored sphalerite and some marcasite. The marcasite, however, is younger than the sphalerite. Consequently mix-crystals of equilibrium composition were not formed.

Because discussions of the conditions existing during ore formation must be based on field work and microscopic study, a further treatment of the specimens one by one is not within the scope of this paper. However, the author is making arrangements of the above mentioned kind for study over a number of Norwegian marmatite ore bodies. The results of this work together with the deduced, pressure corrected, temperatures of formation, will appear in a later issue of this journal.

The Distribution of Minor Elements.

Manganese and cadmium, as earlier discussed, replace zinc in the sphalerite lattice, in the same manner as iron does. The solubility of MnS, CdS respectively in ZnS, of course, is a function of temperature and pressure. The runs conducted with MnS, CdS respectively mixed with ZnS in varying proportions, indicated that both the MnS — ZnS and CdS — ZnS systems are of nearly the same type as the FeS — ZnS system. These runs also supplied evidence as to the maximum solubility of MnS, CdS respectively in ZnS. Without going into details one may conclude, from the above mentioned runs and from the MnS and CdS analyses presented in table XIV, that none of the natural ores analyzed contain nearly as much MnS or CdS as corresponds to equilibrium conditions at the time they were deposited. This of course is not strange because the occurrence of free MnS or CdS formed simultaneously with mix-crystals of ZnS and the mentioned sulfides is, to the author's best knowledge, not on record.

The Mn and Cd distribution in the analyzed samples, therefore, only reflects the local abundances of these elements and says very little about the temperature of formation of the mix-crystals.

The specimens were analyzed on copper and lead as a check on the purity of the picked samples and as a means of correcting the analyses on FeS. The silver reported in the analyses occurs almost entirely in galena.

The analyses on germanium, gallium, nickel, cobalt, and tin do not supply satisfactory information to allow any conclusions to be made as to their concentration in relation to temperature of formation of the mix-crystals in which they occur.

Like manganese and cadmium, the trace elements are hardly ever present in large enough amounts to saturate the ZnS lattice on their sulfides, at the prevailing (P,T) conditions. The amounts present of any one are, therefore, dictated by their local concentrations at the time of mix-crystal formation and very little by the temperature. At any rate, mix-crystals under-saturated on Mn, Cd, Ge, Ga, Ni, Co, and Sn are formed.

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THE TEMPERATURE OF GRANITIZATION IN THE RENDALSVIK AREA, NORTHERN NORWAY

Experimental work by GUNNAR KULLERUD.
Geological discussion by HENRICH NEUMANN.

Rendalsvik is a small village on the peninsula between Holandsfiord and Tjongsfiord south of Bodø and north of Mo i Rana in northern Norway. See map, fig. 1.

A graphite mine near Rendalsvik was worked before and during World War II but was closed down in 1945. It is now planned to re-open and develop the mine on a larger scale. The Geological Survey of Norway (Norges geologiske undersøkelse) has undertaken a thorough investigation of the area comprising field-mapping, geophysical measurements as well as petrological, mineralogical and geochemical studies.

The Holandsfiord—Tjongsfiord peninsula is situated in the central zone of the Caledonides, and in a major axial culmination where the strike of the rocks is locally E—W and thus nearly normal to the prevailing strike (NNE—SSW) of the Caledonides of this region.

The peninsula is made up[1] of granite, gneiss, and metamorphosed sediments in amphibolite facies, see map fig. 2. Because of the folding and subsequent granitization it is not altogether easy to reconstruct the stratigraphical succession, but it is probably, from the base upwards: mica-schists; lime-silicate-gneisses with layers of mica-schists, limestones, and amphibolites; graphite-schists; quartzite; and mica-

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[1] A more detailed account (in English) of the geology of the region by S. SKJESETH and H. SØRENSEN will appear shortly among the publications of the Geological Survey of Norway. We have consulted their manuscript and made use of some of their factual data, maps and sections in the preparation of this paper.

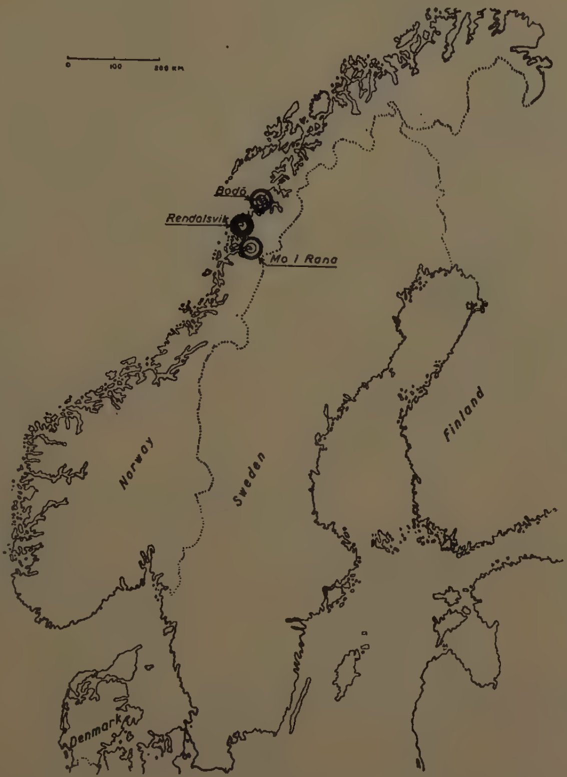

Fig. 1. Map showing situation of Rendalsvik.

schists; i.e. an original sequence of sediments mainly consisting of shales with layers of limestones, black shale and sandstone. The graphite-schist (black shale) is overlain by quartzite (sandstone) with a narrow zone of mica-schist (shale) in between.

The graphite-schist is folded into an anticline or anticlinorium with a steep and compressed southern limb and a more gently dipping and undulating northern limb (see section fig. 3). It is characteristically thickened in the crests of the anticlines.

The granite and granitic gneiss could be more correctly termed granodiorite and granodioritic gneiss. The junction between these rocks and the sediments may be a conformable or unconformable one. In several places there are gradual and continuous transitions from mica-schist via granitic gneiss to fairly homogeneous granite,

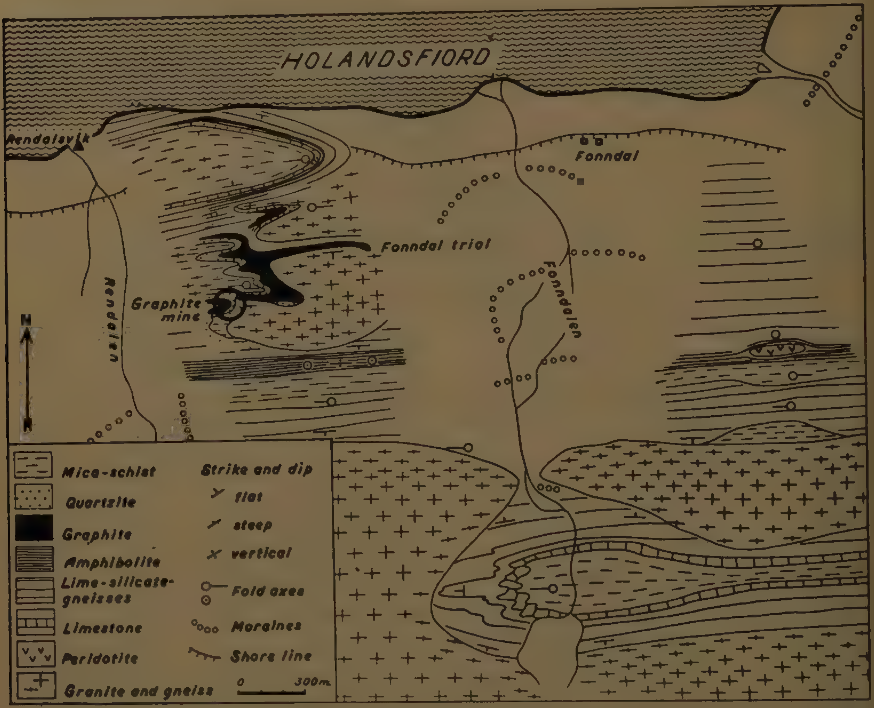

Fig. 2. Geological map of the Rendalsvik area (after S. Skjeseth). Centre of ring denotes place where analyzed sample was collected.

and from impure limestone via lime-silicate-gneiss to biotite-gneiss. The transitional changes can be especially well studied on the bare rock surfaces left uncovered on the withdrawal of the big glaciers Engenbreen and Fonndalsbreen during the last 30 or 40 years. It should also be mentioned that while the granitic rocks may transect the strike of the sediments, the foliation and lineation of the former are always parallel to the strike of the latter. The field evidence suggests a formation by replacement in situ of the pre-existing sediments. A beautiful illustration of this process is found a few metres south of the Fonndal trial (see fig. 2) where the graphite-schist and the overlying narrow band of mica-schist have been granitized: the typical zig-zag, smallscale folding of the mica-schist can be followed across the junction and for about 15 cm into the granite where a very clear "ghost structure" is discernible (see fig. 4).

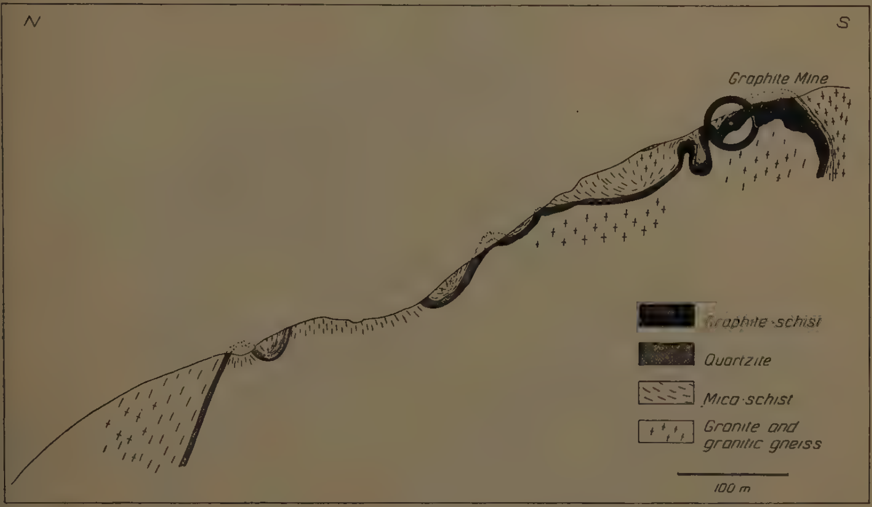

Fig. 3. North-south profile passing through the graphite mine (after S. SKJESETH). Centre of ring denotes place where analyzed sample was collected.

The three major events in the history of the Rendalsvik rocks were: 1) deposition of the sediments in the Caledonian geosyncline, 2) folding (and perhaps thrusting) which, apart from mere details, gave the rocks their present architecture, and 3) granitization, which affected all the sediments, though the graphite-schist least of all. In several places the sediments below and above the graphite-schist have been granitized and the schist itself is now totally enveloped by granite. The reason for the graphite-schist being more resistant to granitization will not be discussed in this paper.

Even if the sediments were mineralogically altered during the period of folding, there can hardly be any doubt that they acquired their present metamorphic state during the granitization period. The metamorphism and the granitization are in fact both manifestations of one and the same process, and the chemical equilibria which can be studied in the schists and gneisses are characteristic for the (P,T) conditions pertaining during the granitization.

The graphite-schist carries sulphides as accessory minerals. They do not originate from the granite, but are primary constituents of the original sediments, as they are evenly distributed in the schist

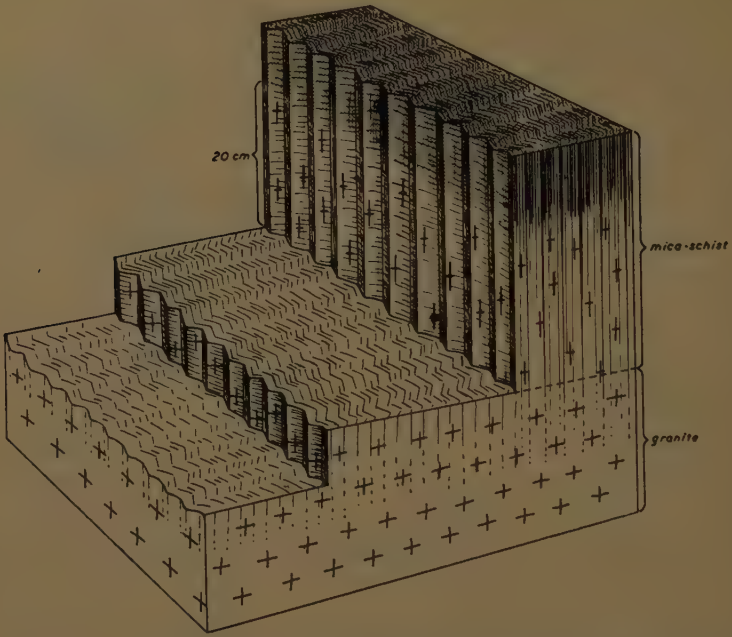

Fig. 4. Block-diagram illustrating the transition from granite to mica-schist south of the Fonndal trial (after S. SKJESETH).

and there are no indications of a metasomatic origin. Hydrothermal veins or veinlets are non-existent.

Pyrrhotite is by far the most abundant sulphide and represents 5—10 per cent of the rock while the cogenetic sphalerite is present in traces only.

When sphalerite crystallizes in the presence of Fe, some Fe will substitute for Zn forming a mixcrystal (Zn, Fe) S. If the amount of Fe be increased, a saturation point will be reached beyond which iron sulphide (as pyrrhotite or pyrite) will form as a separate phase in equilibrium with sphalerite. In the preceding paper of this journal one of us (G.K.) has studied the saturation point as a function of temperature and pressure and shown by experiments how the Fe content of sphalerite in the parageneses sphalerite-pyrrhotite and sphalerite-pyrite can be used as a geological thermometer (and manometer).

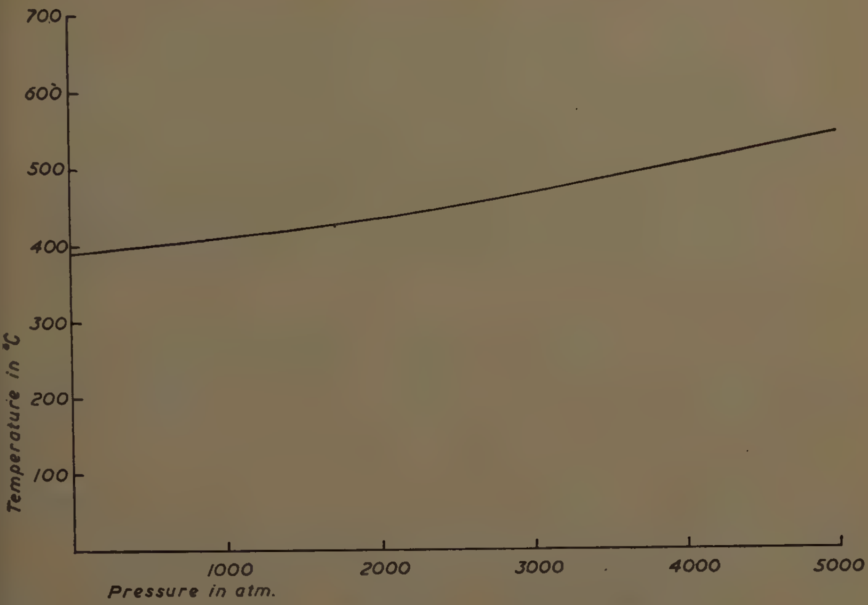

Fig. 5. Curve illustrating pressure-temperature relationship for the formation of the mix-crystals (Zn, Fe) S containing 12.5 mol.% FeS, when formed in equilibrium with pyrite or pyrrhotite.

A representative sample of the graphite-schist was ground to —70 mesh and a sphalerite (marmatite) concentrate obtained by application of heavy liquids. The concentrate was hand-picked under a binocular lens to remove composite grains and the pure sample (not containing more than 1—2 per cent foreign minerals) analyzed for iron and manganese by Miss ERNA CHRISTENSEN in the Chemical Laboratory of the Geological Survey of Norway. The (Zn, Fe) S mix-crystals were found to contain: 12.5 mol % FeS and 4.5 mol % MnS. Unfortunately no analysis was done on cadmium which ,in sulphides of this type is usually present to the extent of 0.1 to 1.0 per cent.

The unit cell length (a_o) of the mix-crystals was found from x-ray powder diagrams to correspond to 12.5 mol % FeS, 4.5 mol % MnS and about 0.8 mol % CdS, thus indicating that exsolution of FeS has not taken place to any measurable extent during the cooling process. The unit cell length versus composition relationship has been discussed in the preceding paper and will not be repeated here. It may only be mentioned that these investigations have shown that the presence of modest amounts of MnS and CdS with the FeS in the ZnS lattice does not influence the FeS solubility to any measurable extent.

When in equilibrium with pyrrhotite or pyrite the mix-crystals containing 12.5 mol % FeS are formed under atmospheric pressure at 390° C (see figure 1 in the preceding paper). The FeS solubility decreases with increasing pressure as shown in table X and figure 9 of the preceding paper.

Fig. 5 below shows the pressure-temperature relationship for mix-crystals containing 12.5 mol % FeS and the sphalerite of the graphite-schist must have been formed under (P,T) conditions corresponding to a point on this curve.

The metamorphism, of course, took place not at one specific temperature but within a temperature range. It is a matter of experience that Fe seems to dissolve more rapidly in the ZnS lattice than it exsolves. If this be the case, the measured temperature of formation will tend to represent a maximum rather than a minimum temperature.

As mentioned above, the schists and gneisses are found in amphibolite facies the temperature range of which has been somewhat differently estimated by various authors.

The sample used for these investigations was collected close to the schist-granite junction. The place is marked with a ring in figs. 2 and 3. Under the regional type of metamorphism which the area has suffered, the temperature of the graphite-schist at the time of equilibrium cannot have been much different from the contemporary temperature within the adjoining granite itself. Granitization must have taken place at the highest temperature which has existed at the schist-granite junction for any length of time and it is concluded that, within narrow limits, the temperature of formation of the sphalerite is identical with the temperature of granitization.

The fundamental data are lacking for an estimation of the average pressure pertaining during the granitization. It is highly improbable that the pressure would have been less than 1000 atm. (corresponding to the hydrostatic pressure at a depth of 3¾ km) and it has probably not been much above 3000 atm. (corresponding to a depth of about 11 km) as there are no signs of phenomena characteristic of the deep roots of mountain chains.

If the average pressure be estimated at 2000 atm. ± 1000 atm., the temperature of granitization in this area was

440° C ± 25° C.

In view of the experimental evidence that no granitic magma can exist below 670° C and accepting the dictum by Read "no magma, no igneous rock", the granite of the Rendalsvik area, in full harmony with the field evidence, is non-igneous.

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NORSK GEOLOGISK TIDSSKRIFT 32

ON THE USE OF ORE MINERALS IN THE INTERPRETATION OF THE STAGE OF METAMORPHISM WITHIN THE AMPHIBOLITE AND SAUSSURITE FACIES

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Abstract. In the present paper the author draws attention to the possibility of using ore minerals, especially the ex-solutions observable in sulphides, in the interpretation of the metamorphic facies of the host rock. In the present paper he considers most common ex-solutions characteristic of the range of temperature corresponding to the amphibolite and to the saussurite facies, and applies microscopic observations in explaining two not very common mineral assiocations of the rocks.

Introduction.

Normal diaphtoretic metamorphism is always accompanied by the incorporation of water. Hence in the metamorphic facies corresponding to lower PT-conditions, the characteristic minerals contain more water (OH) than those belonging to the higher facies. On going from the pyroxene-hornfels facies with the mineral association diopside — hypersthene towards the amphibolite facies with hornblende, water must be added, and the transition temperature between these two facies is, according to ROSENQVIST (1952), that of the expulsion of the water from hornblende — about 750° C at atmospheric pressure.

If water is not available, the change of facies does not occur, or results in assemblages not definitely interpretable. Rosenqvist (1952) called this kind of metamorphism ''dispsenic''. In such cases the determination of the stage of metamorphism is not always easy, and sometimes it is impossible.

In such cases, when the rock also contains ore minerals, the origin (or introduction into rock) of these is controlled by definite conditions, often by a determinable range of temperature under special tectonic circumstances. Usually the first factor determines the possibility of movement of the corresponding ore minerals, the second factor the place where they will be deposited.

The conditions governing the deposition of an ore apply to the silicate minerals of the host rock also; hence the stage of metamorphism during the deposition of the ore will also correspond to the same conditions. Consequently it seems to be possible to find some trends, from the manner of occurrence of ores, which can be used in determining the range of metamorphism in the host rock. Examples of such trends are many kinds of ex-solutions, often observed in the ore minerals. The unmixing of a solid solution takes place by diffusion through the lattice of the solvent substance, and this diffusion through such minerals as oxides and sulphides is so slow that certain textures arising from arrested diffusion persist throughout geological time. (EDWARDS, 1947, p. 56). If there occur in the ore minerals such unmixing phenomena, then they can be used to determine the PT conditions governing the metamorphism of the rock during the deposition of the ore. If ex-solutions do not occur, in many cases assemblages of ore minerals, replacements, alterations, etc. can also be used for determining the PT conditions.

In the present investigation only the amphibolite and saussurite facies will be discussed.

The Amphibolite Facies.

The amphibolite facies of the rocks corresponds to the range of temperature between 400 and 750° C (or the corresponding ''high gneiss facies" of ROSENQVIST (1952) between 400 and 700° C). Among the commonest sulphides occurring in the Finnish ores the following ex-solutions have been studied experimentally; all within the temperature range of the amphibolite facies:

TABLE I.

Host mineral	Ex-soluted mineral	Temperature below which unmixing occurs ° C	Authority
Pyrrhotite	chalcopyrite	600	Hewitt-Schwartz (1937)
Chalcopyrite	sphalerite	550	Buerger (1934)
Chalcopyrite	stannite	500	Ahlfeld (1934)
Pyrrhotite	pentlandite	425—450	Newhouse (1927) Hewitt (1938)
Bornite	chalcopyrite	475	Schwartz (1931)
Chalcopyrite	cubanite	450	Schwartz (1927)

(This table is taken from EDWARDS, 1947, p. 72).

At Puumala, in Southeast Finland, there is a basic cluster embedded in mica schist and characterized by a high content of sulphide minerals together with ilmenite and magnetite as accessories. The main sulphide is pyrrhotite, but pentlandite is also abundant. In addition chalcopyrite occurs. The basic cluster consists in its marginal parts of hornblende, (diopside), plagioclase, and quartz, but the inner part mainly of diopside and basic plagioclase. In the marginal parts the uralitization of diopside has occurred but not in the inner part of the cluster. The rhombic pyroxene does not occur at all, and therefore it was not to be supposed that the inner part of the cluster would correspond to the pyroxene-hornfels facies, but that the mineral association diopside-plagioclase would here correspond to the amphibolite facies also. That there uralitization has not taken place, must be attributed to lack of the water needed for the uralitization of pyroxene. Consequently the assemblage diopside-plagioclase corresponds to the "dipsenic" stage of the amphibolite facies, if the definition of ROSENQVIST (1952) is used. Since the ore minerals are similar both in the marginal and in the inner parts of the cluster, in any case the metamorphism during the deposition of the sulphides must have been similar in all parts of the cluster. Now pentlandite often occurs in the form of flame-shaped bodies in the pyrrhotite, a manner of occurrence which has usually been interpreted as the result of unmixing of the pentlandite from solid solution in pyrrhotite. Such an ex-solution of pentlandite from pyrrhotite occurs at 425—450° C (see the table above) indicating that the metamorphism of the host rock during the deposition of the ore must also have taken place under conditions corresponding to the amphibolite facies; hence from the study of ore minerals new evidence is obtained in favour of the interpretation of the diopside-plagioclase assemblage as corresponding to the dipsenic stage of the amphibolite facies.

The Saussurite Facies.

The saussurite facies of the rocks corresponds to the temperature range between 200 and 400° C. Among the commonest sulphides of the Finish ore minerals the following ex-solutions have been observed corresponding to the temperature conditions of the saussurite facies (se Edwards, 1947, p. 72):

TABLE II.

Host mineral	Ex-soluted mineral	Temperature below which unmixing occurs ° C	Authority
Sphalerite	chalcopyrite	350—400	Borchert (1934)
Chalcopyrite	pyrrhotite	300	Hewitt-Schwartz (1937)
Bornite	chalcosite	175—225	Schwartz (1928)

The last-mentioned ex-solution corresponds already to the tran-itional conditions between the saussurite and the green schist facies.

As an example of the use of the ore minerals in the determination of degree of metamorphism of some lesser known mineral assemblages an instance from Nokia, South Finland, will be taken. There the main rock is normally phyllite, but embedded in this there is a strip of sulphide schist, containing pyrrhotite in abundance but also chalcopyrite and sphalerite as sparse constituents.

The biotite of normal phyllite is quite common, somewhat dark and distinctly pleochroitic with an axial angle of about zero. Sulphide schist contains biotite also, but this is very pale and with axial angle of about zero, exceptionally, however, exceeding upto 16°. This pale biotite is chloritized along its margins, and in extreme cases completely, and then the assemblage corresponds to the saussurite facies. Hence it has been supposed that if pale biotite occurs without chlorite, the assemblage pale biotite — quartz — pyrrhotite belongs to the lowermost degree of the amphibolite facies or to the saussurite facies, and if muscovite is also present, the last-mentioned facies is evident. Among the sulphides, as mentioned in foregoing, chalcopyrite and sphalerite often also occur, and even in the schists containing pale biotite, sphalerite containing unmixed grains of chalcopyrite is common. This phenomenon of ex-solution indicates, as is seen in table II, a temperature of 350—400° C during the deposition of the ore, this being also the temperature governing the metamorphism of the rock during the same period of time; this temperature indicates that at least in the case of Nokia the assemblage pale biotite — quartz — pyrrhotite is that of a transitional stage between the amphibolite and the saussurite facies. Still it may be mentioned, that the sulphide schists of Nokia usually contain some nickel also, and the pyrrhotite separated from sulphide schist of Nokia (with pale biotite) contained 0.06 % nickel (MARMO and MIKKOLA, 1951, p. 36). The flame-shaped bodies of exsolution of pentlandite from pyrrhotite, however, has been observed at Nokia only exceptionally, and then only in schists containing dark biotite, but never in connection with pale biotite. Hence the ex-solution characterized by the temperature range of 425—450° C (see table I) does not occur together with pale biotite, but an ex-solution of sphalerite from chalcopyrite characterized by temperature of 350—400° C (see table II) there occurs.

Unfortunately instances of ex-solutions in common sulphide minerals within the temperature range corresponding to the saussurite facies are very sparse, so limiting the possibilities of interpreting the "sub-stages" of the saussurite facies by using them as thermometers.

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BEMERKNINGER ANGÅENDE VERDIEN AV ELEKTRONØYTRALITETSSKALAEN I DISKUSJONEN AV BINDINGSENERGIEN I SILIKATSTRUKTURER

AV

H. FLOOD.

Den kjemiske affinitetslære har som kjent lenge vært i forgrunnen i mineralogisk og petrografisk forskning. En rekke sentrale spørsmål er med stor fordel blitt behandlet ut fra teoretisk-kjemiske synspunkter. Man ser derfor ofte at nye teoretisk-kjemiske betraktningsmåter allerede etter kort tid taes opp og føres videre av geologer og mineraloger.

Så verdifullt dette forhold i og for seg er, støter man likevel på enkelte tilfelle hvor teorien kan sies å være blitt anvendt på en uforsiktig måte.

Et eksempel på dette gir Paulings elektro-negativitetsskala, som i de siste år av flere forskere er brakt inn i diskusjonen av mineralkjemiske spørsmål [1] [2].

I det etterfølgende skal, etter oppfordring, knyttes et par bemerkninger til disse spørsmål med utgangspunkt i synspunkter som er framholdt i [2].

I to avhandlinger i Journal of Geology påviser H. RAMBERG det forhold at elementene jern og magnesium fordeler seg mellom ko-eksisterende silikat-mineraler på en slik måte at magnesium anrikes i de mer sure, jern i de mer basiske mineraler.

Til denne i og for seg meget interessante påvisning knyttes imidlertid en del teoretiske betraktninger, som kan være egnet til å gi feilaktige forestillinger om de forholdsvis enkle hovedlover som gjelder med hensyn til affinitetsforholdene innen ioneforbindelsenes kjemi.

Da Mg^{++} og Fe^{++} bærer samme ladning og dessuten er lite forskjellige med hensyn til ioneradius, antar RAMBERG at den påviste effekt skyldes årsaker «som ikke er forutsett i den klassiske ioneteori». og han prøver isteden å oppstille en forklaring på grunnlag av de to metallers forskjellige plasering i den Paulingske elektronegativitetsskala.

RAMBERG innleder sine betraktninger med å refere en del reaksjonsvarmer for dannelsen av salter ved omsetning mellom sure og basiske oksyder. Han mener av disse data å kunne stille opp den generelle regel at:

For base-utbytningsreaksjoner (dvs. ione-bytningsreaksjoner) mellom saltpar, vil reaksjonen være eksoterm i retning av dannelsen av salt av den sterkeste syre og det mest elektropositive metalls oksyd.

Som belegg for denne sats anføres reaksjoner som
$Li_2SO_4 + Cs_2CO_3 = Li_2CO_3 + Cs_2SO_4 \quad \triangle H = -19.85$ kcal.

Dette er egnet til å gi inntrykk av at drivkraften for reaksjonen skulle være at det mest elektropositive metallion Cs^+ søker å forene seg med det sterkeste syreradikal — dvs. danne sulfat.

At denne regel har høyst tvilsom verdi som generell sats, framgår allerede av den korresponderende reaksjon.
$2\,LiCl + Cs_2CO_3 = Li_2CO_3 + 2\,CsCl$ — hvor $\triangle H \sim +25$ kcal,
altså stikk motsatt hva regelen forlanger;
eller at

$$LiCl + KF = KCl + LiF$$

$\triangle H = +17$ kcal, til tross for at HCl er en sterkere syre enn HF.

Årsaken til dette er det ikke vanskelig å se, og jeg skal om et øyeblikk komme tilbake til den i en litt videre sammenheng.

For nå å få elektronegativitetsskalaen inn i billedet, postulerer RAMBERG at surstoffet vil opptre som om dets elektronegativitet avhenger av dets binding, og det på en slik måte at den forsterkes i rekkefølgen:

orto — pyro — meta osv.

På denne måte kommer han fram til den oppfatning at det som bevirker magnesiumets anriking på bekostning av jernet i de sure mineraler, er magnesium-ionets tendens som det mest positive i skalaen av de to til å få legge beslag på de mest elektronegative surstoffioner.

Det er denne konklusjon som er egnet til å gi et helt forkjært bilde av forholdene. — Først kunne det imidlertid være fristende å si et par ord om det formålstjenlige i å trekke elektronegativitetsskalaen inn i diskusjonen av affinitetsforholdene av ioneforbindelser.

Den teoretiske bakgrunn for elektronegativitetsskalaen kan sammenfattes slik:

I en normal kovalent binding mellom atomene A og B i et molekyl A B settes bindingsenergien D lik det aritmetiske (evt. geometriske) middel av de to bindingsenergier A—A og B—B.

$$D_{AB} = \frac{1}{2} (D_{AA} + D_{BB}).$$

Hvis A og B avviker med hensyn til elektroneaffinitet, kommer i tillegg et resonans bidrag Δ som skyldes ionekarakteren i bindingen:

$$D_{AB} = \frac{1}{2} (D_{AA} + D_{BB}) + \Delta$$

hvor Δ er desto større jo mer de to atomer atskiller seg fra hverandre i elektroneaffinitet.

Hvis forskjellen ikke er for stor, kan Δ tilnærmet uttrykkes av formelen

$$\Delta = 23.06 \ (x_A - x_B)^2$$

hvor x er en empirisk konstant for hvert atom, betegnet dets «electronegativity value». Disse x-verdier er det som sammenfattes i elektronegativitetsskalaen. Verdien av denne formel skal ikke bestrides, sålenge det gjelder bindinger av vesentlig kovalent karakter.

Men det er helt utenfor teoriens gyldighetsområde å bringe den til anvendelse på forbindelser hvor ionekarakteren er den dominerende. F.eks. Δ H for reaksjonen

$$\frac{1}{2} Cs_{2(g)} + \frac{1}{2} F_{2(g)} = CsF_{(g)}$$

beregnes etter elektronegativitetsskalaen til ∼ —280 kcal, mens den eksperimentelt finnes lik ∼ —110 kcal. Altså et fullstendig ubrukelig resultat.

Grunnlaget for en generell sats om at ionebytningsreaksjoner

Det fremgår forøvrig at forskjellen mellom to atomer i elektronegativitetsskalaen kun forteller om Δ-bidraget i bindingsenergien, men intet om bindingsenergien i sin helhet.

mellom to saltpar forløper eksotermt i retning av dannelse av salt av det mest elektropositive og det mest elektronegative element, er derfor mer enn spinkelt. Hvor galt dette kan føre, framgår f.eks. av reaksjonen

$$CsF + LiJ = LiF + CsJ, \ \Delta H = —33 kcal.$$

Forbindelsen mellom det mest elektropositive element Cs og det mest elektronegative F spaltes opp under omsetning med LiJ under frigjøring av en betraktelig varmemengde.

Når de problemer det her gjelder ikke har vært fremme i den uorganiske kjemi de siste årtier, er årsaken rett og slett den at for den klassiske elektrostatiske teori reiser ikke forståelsen av dette noen vanskelighet.

Reaksjonsvarmen er et uttrykk for at summen av gitterenergiene på høyre side er større (mer negativ) enn på venstre. Det er LiF som har den største gitterenergi. Reaksjonsvarmen manifesterer derfor ganske enkelt kravet fra ionene med de mindre radier om å bli bundet sammen, da denne kombinasjon vil gi den laveste elektrostatiske energi.

[Det bør samtidig bemerkes at graden av kovalens i en H—X binding på ingen måte (som man kunne slutte av Rambergs anførsler) står i noen enkel relasjon til syrestyrken (protonavspaltningstendensen) sml. f.eks.

$$HF < HCl < \text{etc.}$$

$$H_2O < H_2S < \text{etc.}$$

$$NH_4^+ < PH_4^+ < \text{etc.}$$

hvor syrestyrken *vokser* med graden av ko-valens.

Også disse forhold forklares tvangsfritt av den enkle elektrostatiske teori ved at $r_F \text{—} > r_{Cl} \text{—}$ osv. Den elektrostatiske teori for styrken av oksysyrer er i sin gamle og velkjente form fullt tilstrekkelig til å forklare de syreegenskaper det her spørres om. (Sml. f.eks. at den umiddelbart gjør det klart hvorfor forskjellen mellom K_1 og K_2 er så meget større for orto- enn for pyrosyrer)].

Det skulle av dette framgå at det er gitterenergetiske betraktninger som må legges til grunn for diskusjonen av affinitetsforholdene for ioneforbindelser. Det er også mulig ut fra meget enkle gitterenergetiske betraktningsmåter å forklare såvel Mg—Fe-fordelingen, som syrestyrken av oksysyrer, de problemer som Ramberg fortrinnsvis beskjeftiger seg med i sin avhandling.

Vi skal innskrenke oss til å betrakte Mg—Fe-fordelingen, da det her synes som om Rambergs forklaringsforsøk kommer i direkte konflikt med de gitterenergetiske betraktninger.

Det som skiller Mg^{++} og Fe^{++} med hensyn til kjemiske egenskaper kan, iallfall så lenge det kun dreier seg om kvalitative vurderinger, enklest forklares ut fra ionedeformasjons-betraktninger.

Som framholdt av FAJANS, GOLDSCHMIDT m.fl., er ved siden av ladning og radius, ionepolariserbarheten hovedfaktor for bindingsenergien av ioneforbindelser. — Disse synspunkter er med stort hell anvendt til forklaring av en lang rekke spørsmål innen ioneforbindelsenes kjemi, sml. f.eks. avhandlinger fra de senere år av W. WEYL og medarbeidere.

Edelgassionet Mg^{++} har et mer stabilt og stivt, mindre polarisertbart elektroneskal (og dermed også mindre polariserende) enn overgangselementionet Fe^{++}. Denne polariserbarhet bevirker et tillegg i ionebindingsenergi, utover den energi som betinges av ladning og radius. Sammenliknes derfor bindingsenergien for Fe^{++} og Mg^{++} til et og samme anion, vil Fe^{++} danne de sterkere bindinger. Forskjellen vil gjøre seg desto mer gjeldende jo mer polariserbart anionet er.

Surstoff-ionet er et meget sterkt polariserbart anion. Sammenlikner man nå surstoff-syrene i rekken orto-pyro-meta, forståes uten videre at surstoff-ionet må være sterkest polarisert av syrens centralatom i meta, svakest i orto-syren. Men derav følger igjen at surstoffionets (resterende) polariserbarhet er sterkest i orto og svakest i meta-syren.

Jern- og magnesiumsfordelingen forklares da tvangsfritt som en følge av at det sterkest polariserende (og polariserbare) kation Fe^{++} fortrinnsvis vil bindes til det sterkest polariserbare anion — ortosurstoffionet.

RAMBERGS betraktninger er egnet til å gi det stikk motsatte inntrykk — at det er Mg-ionets tendens til å bindes til pyro- (eller meta)-surstoffionet som er det avgjørende.

Forklaringen av virkningen av substitusjon av Si med Al gir seg av seg selv. Si polariserer surstoffionet sterkere enn Al. Erstatning av Si med Al fører derfor til svakere polariserte, dvs. sterkere polariserbare surstoffioner, hvilket igjen medfører binding av jern på bekostning av magnesium.

Til slutt bør kanskje for tydelighets skyld tilføyes at det som her er anført ikke må oppfattes slik at de enkle elektrostatiske forestillinger er istand til å forklare fullt ut alle affinitetsproblemer man støter på i ioneforbindelsenes kjemi. Tvert imot er det lett å peke på forhold som utvilsomt krever kompletterende antakelser. (Sml.

f.eks. de rettete valenser, som også setter sitt tydelige preg på ionekrystallkjemien).

Det skulle allikevel være nokså klart at den enkle elektrostatiske ionegitterteori opererer med en modell som representerer en bedre første tilnærmelse av de bindingsenergetiske forhold i ionekrystaller enn teorier på grunnlag av elektronegativitetsskalaen, som tar sitt utgangspunkt i den rent kovalente binding.

English Summary.

During the last years attempts have been made to relate mineralogic ionic affinity problems with Paulings electronegativity scale of the elements.

It is pointed out that the electronegativity scale cannot be applied to typical ionic bonds.

A better and more comprehensive approach to the problems in question is represented by the simple ionic model.

[1] J. W. Gruner. American Mineralogist, Vol. 35, pp. 137—148, 1950.
[2] H. Ramberg. The Journal of Geology, Vol. 59, pp. 193—110, 1951. Vol. 60, pp. 331—355, 1952.

Manuskript mottatt 26. mars 1953

LITT OM TROLLHEIMEN UNDER SISTE ISTID

AV

ARNE GRØNLIE.

Det har i de siste åra vært ført en livlig diskusjon i Naturen om det var isfri områder i Trollheimen under siste istid. Diskusjonen ble reist av NILS A. SØRENSEN som i mars 1949 hadde en artikkel med tittel: «Gjevilvasskammene — nunatakker i Trollheimens midte ?». SØRENSEN peker her på at det på Hemre Gjelvilvasskam i ei høgd på ca. 1400 m o.h. er 3 steinfri jordlag som han meiner er fra siste interglasialtid. På toppen av Midtre Gjevilvasskam (1640 m o.h.) er et liknende jordlag. Vi skal ha å gjøre med dannelser av noenlunde samme slag som ROLF NORDHAGEN har påvist bl.a. på Sørøya (NORDHAGEN 1945 og seinere).[1] Han konkluderer med at dersom forvitringsjordrestene i Gjevilvasskammene er interglasiale, kunne siste istids kåpebre i disse områda ikke ha nådd høgre opp enn til et nivå som i dag ligger ca. 1350 m o.h. Dette meiner han ikke står i strid med resultata av den kvartærologiske forsking på Møre og i Sør-Trøndelag. For å vise dette finner SØRENSEN gradienten for breen fra Gjevilvasskammene til Ørlandsstadiet og til UNDÅS' Rundestadium.[2] Dersom han forutsetter at Ørlandsstadiet representerer yttergrensa for siste istids ismasser, får han en gradient som varierer mellom 1/70 og 1/100. (Han rekner da gradienten langs de naturlige passasjer og daler som isen måtte følge). Dette meiner han er for store verdier. Dersom Rundestadiet representerer den ytre grensa, får han en gradient som varierer mellom 1/120 og 1/160. Disse verdier meiner han er så noenlunde i samsvar med hva en måtte vente. SØRENSEN peker også på botaniske faktorer som synes å tyde på at det var isfri områder i Kammene der visse planter kunne «overvintre». Særlig vekt legger han på å ha funnet *Arenaria norvegica* på Hemre Gjevilvasskam.

SØRENSEN har i de seinere år fortsatt sine kvartærgeologiske undersøkelser i Trollheimen og har dessuten utvida undersøkelsene sine nordvestover til havet, idet han har forfølgt et snitt fra Gjevilvasskammene og ut til Tustna-Stabben. Resultata er ikke publisert. Men SØRENSEN har vært så elskverdig å sende meg et resymé av undersøkelsene. Av dette går det fram at han meiner at det i Trollheimen eksisterer to «generasjoner» flyttblokker, en med blokker som er mye forvitra og som synes å ligge på ei overflate som er sterkt tært av tidens tann, og en med blokker som er lite medtatt av forvitring og som ligger på ei «frisk» overflate. Den første blokktypen ligger høgst i terrenget og stammer fra ei eldre nedising, mens den siste skulle være fra siste istid. Ikke engang den mektigste av de to istidene hadde ei isoverflate som strakte seg opp over de høgste toppene. På grunnlag av studium av blokkene og den overflata de ligger på, skulle det etter SØRENSEN gå an å fiksere den øvre grensa for ismassene fra de to istidene, men det må nøyere undersøkelser til. Han viser til ei rekke observasjoner av de to flyttblokk- og forvitringstyper. Her skal bare refereres at når det gjelder Hemre Gjevilvasskam meiner han at flyttblokkene på den sørlige «snute» (karthøgd 1329 m o.h.) er heilt friske og ligger på bart fjell. Nord for fjellhakka (se seinere) der kammen har ei høgd på 1410 m o.h., skal derimot blokkene være sterkt forvitra. Mellom disse to nivåer må da den øvre grensa for den siste istids ismasser ligge. Dette samsvarer med den høgda som han oppgir for dette nivå i sin artikkel i Naturen som er referert ovafor.

Hva snittet mot nordvest angår, legger han særlig vekt på undersøkelser som er foretatt sommeren 1951 på Tustna-Stabben. Her kan påvises ei veldig isskuring opp til 490 m o.h. som viskes ut videre oppover. I ei høgd på 700 m o.h. er praktfulle jorddekker som han meiner viser at isen aldri har gått så høgt.

HANS HOLTEDAHL, som har foretatt geologiske undersøkelser i det nordvestre hjørne av rektangelkartet Oppdal (mellom Storlidalen, Sunndalen og den vestre avgrensing av kartbladet), meiner at dette området har vært totalnedist under siste istid (1949, 1950).[3] Han bygger dette bl.a. på at en skulle vente å finne en markert skilnad i forvitring mellom et nedre område som har vært nedist og et øvre som ikke har vært det, noe som han ikke har kunnet påvise. Han har videre funnet nord-sørgående skuringsstriper på Gråfjellet og

Sandåfjellet i ei høgd på 1550 m o.h. og flyttblokker og aust-vestgående skuringsstriper på fjellpartiet Gresslihø i ei høgd av ca. 1400 m. Endelig har han på vestskråningen av Otthø funnet nord-sørgående skuringsstriper og flyttblokker i 1300—1400 m høgd. Han trur ikke at SØRENSEN kan ha rett i at det maksimale isnivået under siste istid ikke rakk høgre enn til 1350 m o.h. i Gjevilvasskammene. I det området han har undersøkt, er det en serie botner med skarpe former og kvasse egger som ikke kan ha motstått virkninga av en innlandsis. Denne botnerosjonen skal ha foregått etter totalnedisinga ved nivasjon (snøerosjon). Skjerping av botnene skal ha skjedd ved frysings- og tiningsprosesser der vatn fra fonner og snø som smelter trenger gjennom bergarten. Ved solifluksjon og av rennende vatn skal en del av forvitringsmaterialet være transportert vekk. Han peker til slutt på at forvitringa av bergartene er avhengig av ei rekke forskjellige faktorer og at en skal være varsom med å slutte for mye av slike jordlag som dem SØRENSEN omtaler fra Trollheimen.

PER HOLMSEN var sommeren 1950 en tur til jordlaga på Hemre Gjevilvasskam. Han peker i en artikkel i Naturen (april 1951)[4] på at SØRENSEN ikke nevner at jordlaga er knytta til de lause glimmerskifrene i trakten. Han meiner videre å ha påvist at de 25—90 cm tjukke jordlaga ligger oppå ei overflate av blokkmateriale som er danna ved frostsprenging. Grensa mot blokkunderlaget skal være skarp. Da jordlaga viser skråskikting, drar han den slutning at vi har å gjøre med vindsortert materiale. Jordlaga ligger så å si heilt fram mot et bratt stup. Når vinden presses mot dette, danner det seg en hvirvel eller et lunt område i le av stupkanten.

HOLMSEN meiner videre at en ikke kan vente å finne flyttblokker på de høgste toppene. I områder der landskapet er oppskåret av daler, vil farten på isen tilta, og flyttblokkene vil synke i terrenget. Videre peker han på det faktum at toppene lenge lå over firngrensa og at vi derfor ikke kunne vente akkumulasjon.

HOLMSEN har funnet flyttblokker til 1500 m høgd i Storhornet sør for Gjevilvatnet, og, som HANS HOLTEDAHL, tviler han derfor på at det var isfri områder i Gjelvilsvasskammene under siste istid.

Videre meiner HOLMSEN at skal en kunne slutte noe om det har vært isfri områder eller ikke ved å studere forvitringsgraden, må en først få en objektiv metode til å bestemme alderen på forvitringsmateriale.

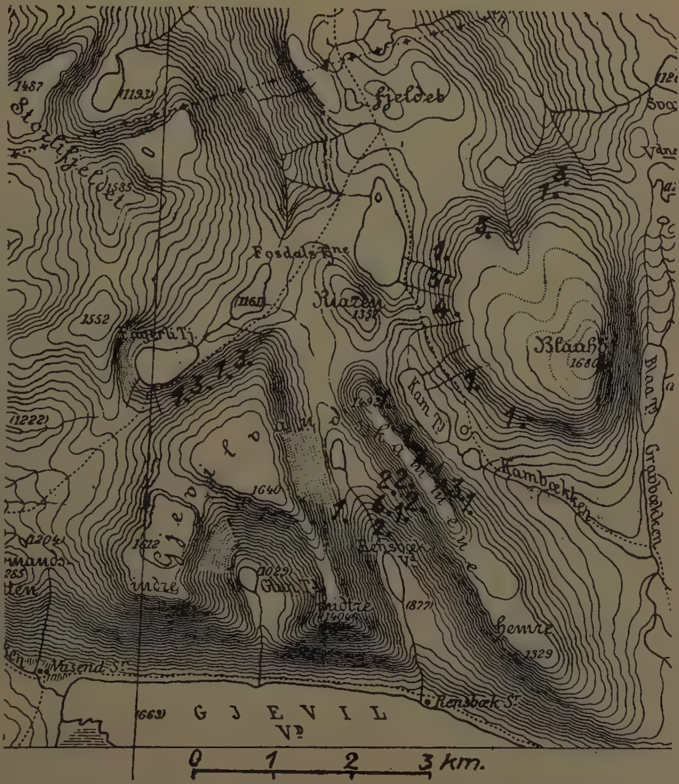

Fig. 1. Utsnitt av rektangelkartet Trollhetta (42 C). Finnestedene for noen av de viktigste «glaciale overvintrere» er avmerkt med tall: 1) *Papayer radicatum,* 2) *Arenaria norvegiac,* 3) *Luzula artica,* 4) *Sagina caespitosa,* 5) *Taraxacum dovrense,* 6) *Nigritella nigra.*

(Kartgrunnlaget er offentliggjort med løyve fra Norges Geografiske Oppmåling.)

OLAV GJÆREVOLL har i flere år arbeidd med floraen i Trollheimen og har gjort interessante nyoppdagelser.[5] Særlig i og omkring Gjevilvasskammene har han heilt til det siste funnet arter som er nye for Trollheimen og som ved sin bi- eller unisentriske karakter, eller på annen måte, tyder på at vi i Møre- og Trøndelagtraktene har hatt isfri områder der plantene har kunnet «overvintre» under siste istid. Da de siste funna ikke har vært publisert før, har GJÆREVOLL vært så elskverdig å sende meg et kart over Gjevilvasskammene med omgivelser, der han har tegna inn finnestedene for noen av de plantene, funnet av han sjøl og andre, som han i dag oppfatter som de viktigste indikatorer på at ei slik overvintring har funnet sted (fig. 1). På kartet er finnestedene for følgende arter avmerkt: *Papaver radicatum, Arenaria norvegica, Luzula arctica, Sagina caespitosa, Taraxacum dovrense* og *Nigritella nigra.*

For ikke å gjøre kartet for overlesst har han tatt med bare disse artene, men peker samstundes på at det i det samme området vokser planter som *Minuartia rubella, Euphrasia lapponica, Arabis patraea, Draba lactea,* alle i betydelige mengder, og til slutt den uhyre viktige unisentriske *Artemisia norvegica* som er en av de vanligste plantene i området.

Den sterke konsentrasjon av de sjeldne plantene i og omkring Gjevilvasskammene, får GJÆREVOLL (som SØRENSEN) til å anta som sannsynlig at Gjevilvasskammene har vært nunatakker under siste istid og at plantene har «overvintret» der.

EILIF DAHL kommer i en artikkel i Naturen[6] ikke direkte inn på tilhøva i Trollheimen, men det er verdt å referere den her, da den diskuterer problemer som er nær knytta til dem vi her har tatt for oss.

Ut fra botaniske indikasjoner slutter DAHL at det må ha vært isfri områder på ytre Møre og i ytre Troms under siste istid. Her må ha vært refugier der plantene kunne «overvintre». Grunnen til at nettopp disse områda har vært refugier, meiner han er at Egga ligger nærmere land her enn ellers langs kysten.

Breen kan ikke ha gått lenger ut enn til Egga. Kalvingsfronten her kunne ikke være videre høg om en sammenlikner med kalvende breer i dag. Gradienten for breie og mektige breer som noenlunde uforstyrra av fjell går ut i havet, er ca. 1/200. Sjøl om han rekner med en gradient på 1/100, vil det bli en heil serie med kystfjell som stikker opp som nunatakker. En ser også i nåtida at høge fjellområder som ligger ved havet, danner en barriere for isstraumen slik at en får isfri områder i le heilt ned til havnivået. DAHL meiner at det må ha vært slik på kysten hos oss.

Refugien inndeles av DAHL i to hovedtyper, kystfjelltypen og tundratypen. Den første har rimeligvis hatt et forholdsvis fuktig klima med mye nedbør, mens den andre typen skyldes et kontinentalt klima med liten nedbør. Kystfjelltypen deles igjen i to undertyper: skandinavisk og antarktisk type, den første karakterisert ved at firn-

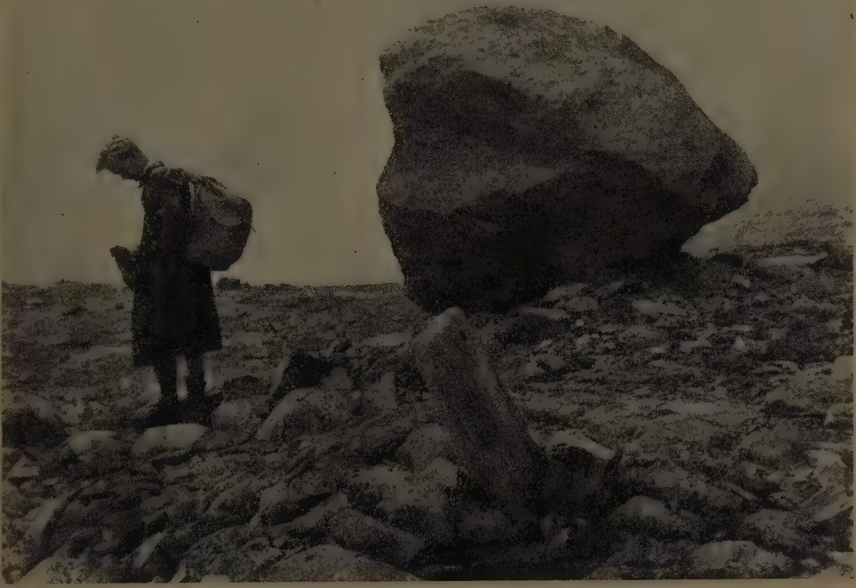

Fig. 2. De høgstliggende flyttblokkene i Trollhetta ligger opptil 1300 m o.h. De er lite forvitra.

grensa ligger over isfritt område, og den andre ved at firngrensa ligger lågere enn det isfri land. På kysten hos oss skal vi ha hatt refugier av den skandinaviske type.

Egne undersøkelser.

Sia 1946 har jeg foretatt en del kvartærgeologiske undersøkelser i Trollheimen. Disse har ikke vært så systematiske som jeg hadde kunnet ønske det. Men da resultata av undersøkelsene kanskje kan ha noen betydning i den diskusjonen som pågår, vil jeg her referere en del iakttakelser som knytter seg til området fra Gjevilvatnet og nord-nordvestover inn i den sentrale Trollheimen.

Jeg har et par ganger gått over Trollhetta. På toppene var det ikke mulig å finne flyttblokker. Under nedstiginga til Folldalen la jeg i ca. 1300 m høgd merke til ei grense mellom et øvre område uten og et nedre område med tallrike flyttblokker. Sjøl de flyttblokkene som lå høgst, viste seg å være svært lite forvitra (fig. 2). Dette grensenivået ligger noe nedafor den formasjonsgrensa som OLAF HOLTEDAHL (1939)[7] har påvist mellom toppdekket av kvartsitt (sparagmitt) og en rødaktig granitt som ligger under, og som trulig er av grunnfjellsalder. En må kunne gå ut fra som sikkert at det ikke kan ha eksistert flyttblokker fra siste istid, som nå er kommet bort som følge av forvitring, over 1300 m nivået. Det er jo ingen grunn til å tru at vi akkurat i 1300 m høgd skulle ha et markert skille i den fart som forvitringa foregår med.

Jeg har også vært oppe på Snota. Her er ikke tilhøva fullt så klare. Det er bl.a. litt vanskelig å avgjøre sikkert hva som er flyttblokker og hva som er stedeget materiale. Særlig gjelder dette noen gneisbergarter. Men jeg kom da også her til det resultat at ingen sikker flyttblokk kunne påvises over et nivå på vel 1300 m o.h.

På Hemre Gjevilvasskam har jeg lett meget omhyggelig etter flyttblokker, særlig langs den søraustre delen. Går vi her fra søraust mot nordvest, viser det seg at undergrunnen til å begynne med består av kvartsitt (sparagmitt) som på Trollhetta. Denne delen av kammen går opp til ca. 1460 m o.h. Her skifter bergarten plutselig over til lause (kambrosiluriske) skifre. Videre nordvestover er det vanskelig å komme fordi erosjonen har skapt noen «hakk» i skiferbergartene parallelt med lagdelinga. Og da lagstillinga er steil, blir kantene bratte. Et stykke søraust for formasjonsgrensa fant jeg to flyttblokker. Den øvste var en øyegneis som lå 1421 m o.h. Blokkene var lite forvitra (fig. 3).

De to gangene jeg har vært oppe på Midtre Gjevilvasskam, har jeg ikke kunnet finne flyttblokker så høgt som i Hemre kam. Jeg trur derfor at det neppe finnes flyttblokker høgre enn ca. 1420 m i Gjevilvasskammene.

Jeg har ikke kunnet påvise to generasjoner av flyttblokker eller noen forskjell i forvitringsgraden av fjelloverflata som ikke får sin naturlige forklaring ved å ta hensyn til at de forskjellige bergarter yter varierende motstand mot forvitring. Jeg har ikke spesielt undersøkt en slik mulighet, men jeg tviler på at SØRENSEN kan ha rett i sin antakelse.

Isskuring har jeg ikke kunnet påvise på noen av de høge toppene

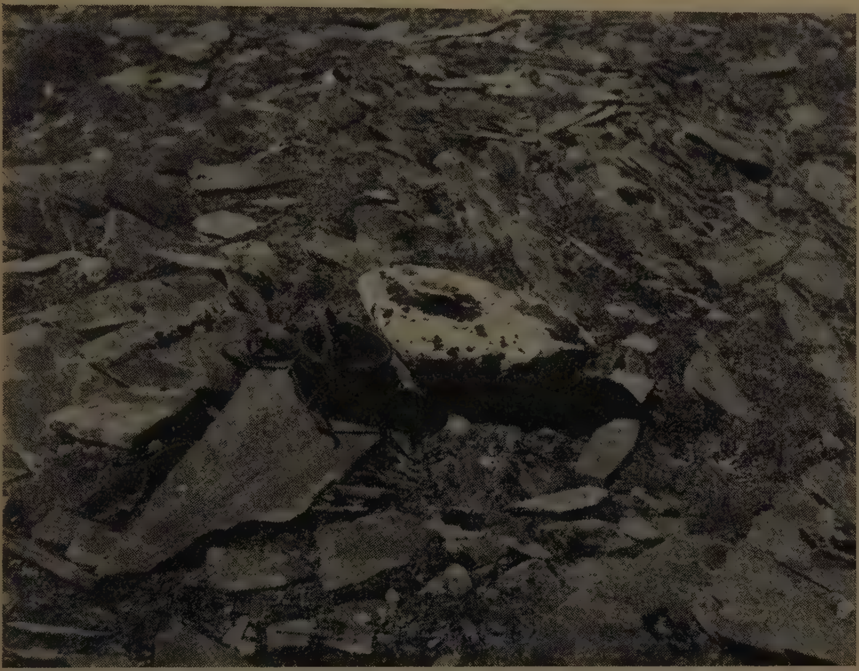

Fig. 3. Den høgstliggende flyttblokk i Hemre kam, 1421 m o.h.

i Trollheimen som jeg har vært oppå. (Her skylder jeg imidlertid å gjøre merksam på at tilhøva sjelden ligger slik til rette at en skulle vente å finne isskuring, sjøl om toppene hadde vært nedist under siste istid. I toppene har vi nemlig i stor grad enten kvartsitter, som svært lett spalter opp, eller lause skifre, som lett forvitrer.) I lågere nivåer har jeg imidlertid flere steder påvist vakker isskuring. I skråninga fra Trollhetta og ned mot Folldalen er i ei høgd av 1184 m o.h. pen isskuring og bogeforma innhakk i berget (fig. 4) som viser at en isstraum har bevegd seg i nord-nordvestlig retning, dvs. nedover Folldalen. På vestsida av Slettådalen, ikke langt fra Slettåfoss, er isskuring og rundsva (med støtside mot søraust) som viser at vi her har hatt rørsle mot nordvest; det vil i dette tilfelle si nedover Slettådalen. På Mellomfjell har vi, like nord-nordvest for det store Fossdalstjernet, bogeforma hakk som viser rørsle i samme lei, denne gang nedover Fossådalen. I Nøstådalen har vi rett nord for Neådalssnota

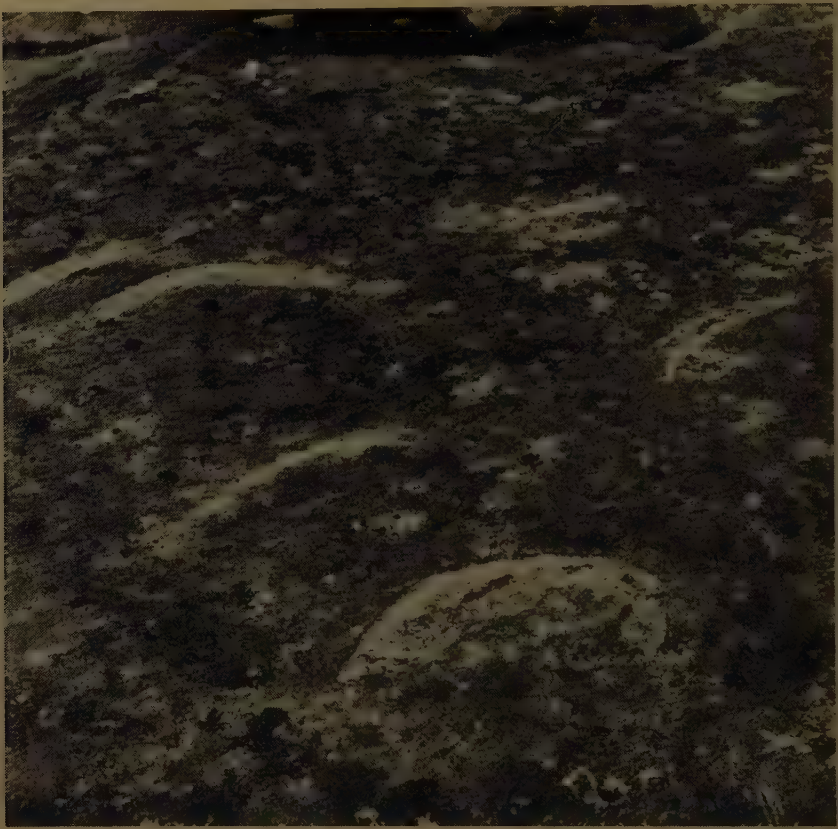

Fig. 4. Bogeforma innhakk i fjellet 1184 m o.h. i Trollhetta. Hakka viser isrørsla mot nordvest, nedover Folldalen.

isskuring mot vest, altså langs Nøstådalen. Like aust for Oppdal stasjon peker ei vakker isskuring mot nordvest.

Når det gjelder jordlaga på Hemre og Midtre Gjevilvasskam, så har jeg undersøkt disse nøyere ved å grave et snitt begge steder. På Hemre kam ble snittet lagt i den ytre kanten av det høgstliggende jordlaget (1428 m o.h.) (fig. 5 og 6). Totallengda på snittet var 130 cm. Det var ikke mulig å påvise noen markert grense (som Holmsen omtaler) i heile snittet. De øvste 90 cm så ut til å være heilt vanlig forvitringsjord, mens de nedste 40 cm var en mellomting mellom for-

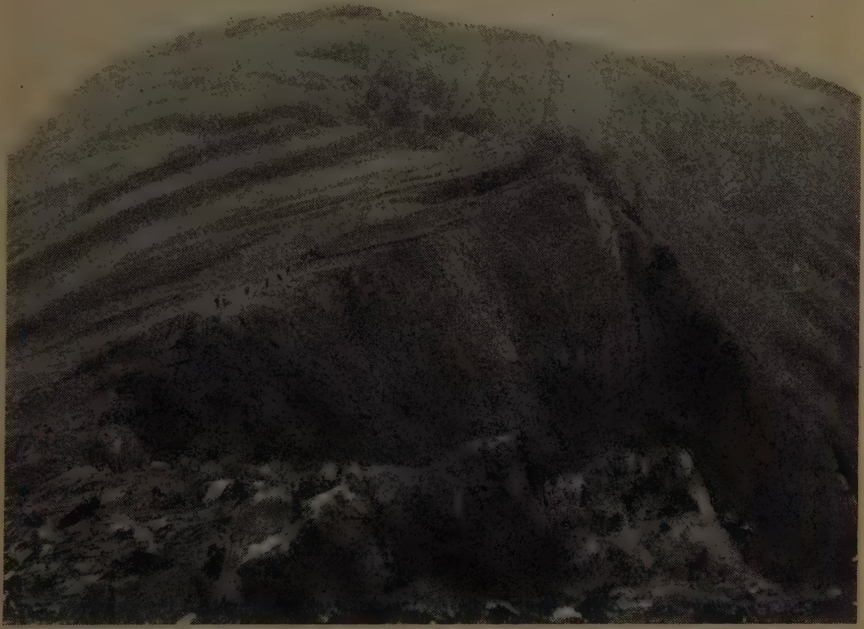

Fig. 5. Her ser vi tydelig jordlaga på Hemre kam. Mest iøynefallende er laget nærmest menneskene.

vitringsjord og berg. Men det var heller ikke her vanskelig å grave. En kunne sikkert ha spadd seg enda langt nedover. Gjennom heile det 90 cm tjukke jordlaget lå hist og her ei og anna lita steinflis som forvitringa ikke har fått bukt med. Det kan neppe være tvil om at vi her har å gjøre med ei kraftig forvitring *in situ*. Vinden får litt tak langs kantene av de oppstikkende jordlaga, så vegetasjonen undergraves, og en får inntrykk av at jordlaga etter hvert langsomt vil forsvinne.

På Midtre kam ble det gravd ei 63 cm djup grøft (fig. 7) noen meter fra jordlaget (ca. 1640 m o.h.). På overflata lå et lag med mindre heller som først ble fjerna. Under var det vanlig forvitringsjord av en litt grovere karakter enn på Hemre kam. Heller ikke her var det

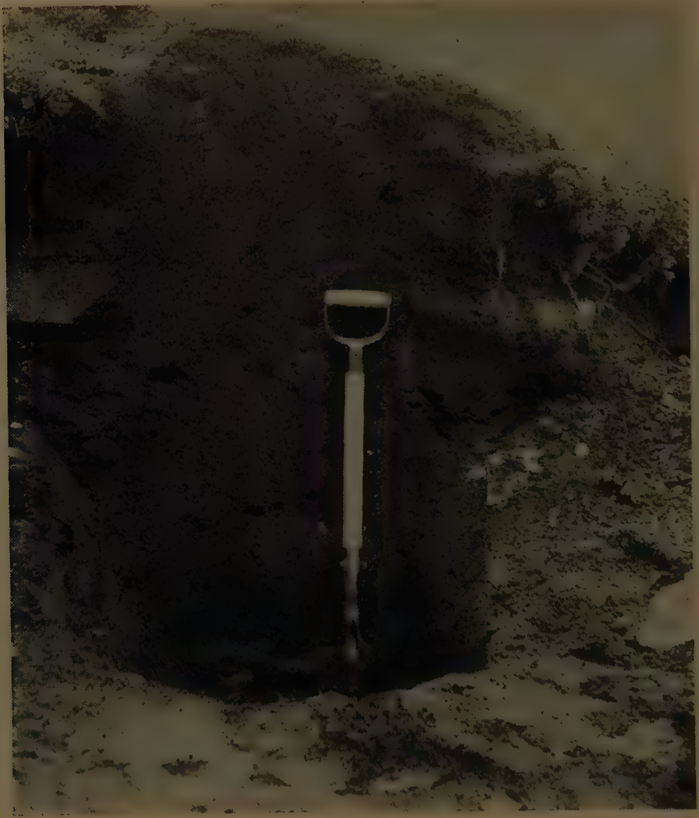

Fig. 6. Snitt i kanten av det øvste jordlaget på Hemre kam. En ser ingen markert grense i snittet. Det heile ser ut til å være forvitringsjord.

noen skarp grense mot berget under. Hva sjølve jordlaget angår, har det samme karakter som laga på Hemre kam (fig. 8).

Jeg kan således ikke være enig med PER HOLMSEN i at vi utelukkende har å gjøre med jord som er blåst sammen. Ei anna sak er at en del av materialet i de øvste laga nok er flytta en del av vinden. Men den alt overveiende del av jorda ligger sikkert på sin opphavlige plass. Det er riktig at alle jordlaga ligger i et «lunt» område ved kanten av et stup, og det er vel nettopp dette som har gjort at jorda ikke er blåst bort.

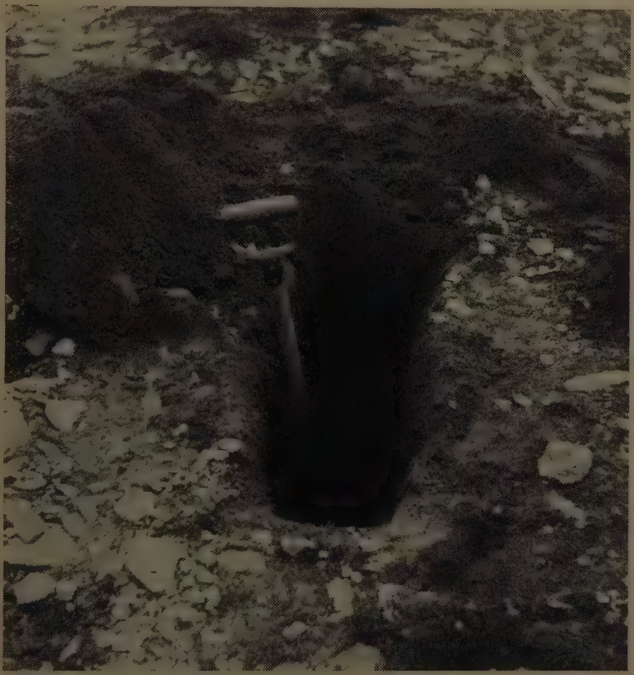

Fig. 7. Her er det blitt gravd ei lita grøft like i nærheten av jordlaget på Midtre Gjevilvasskam. Vi har også her å gjøre med forvitringsjord som går jamt over i berget under.

Diskusjon.

Vi kan slå fast at i den del av Trollheimen som er omtalt i dette arbeid, har det ikke vært påvist flyttblokker eller skuringsfenomener over et nivå som i Hornet ligger ca. 1500 m o.h., i Gråfjellet og Sandåfjellet ca. 1550 m, i fjellpartiet Gresslihø ca. 1400 m, ved Otthø 13—1400 m, i Hemre Gjevilvasskam ca. 1420 m og i Trollhetta ca. 1300 m o.h. Ingen av skuringsmerkene ligger over 1200 m o.h., og alle — bortsett fra den isskuringa som HANS HOLTEDAHL har påvist på Gråfjellet og Sandåfjellet — peker langs dalfører eller andre passasjer gjennom fjellmassivet i relativt moderat høgd. En får absolutt det inntrykk at ismassene ikke har vært mektigere enn at de har måttet innrette si rørsleretning etter relieffet og bare har kunnet trenge seg fram langs daler, pass og skar. Som nevnt danner isskuringa

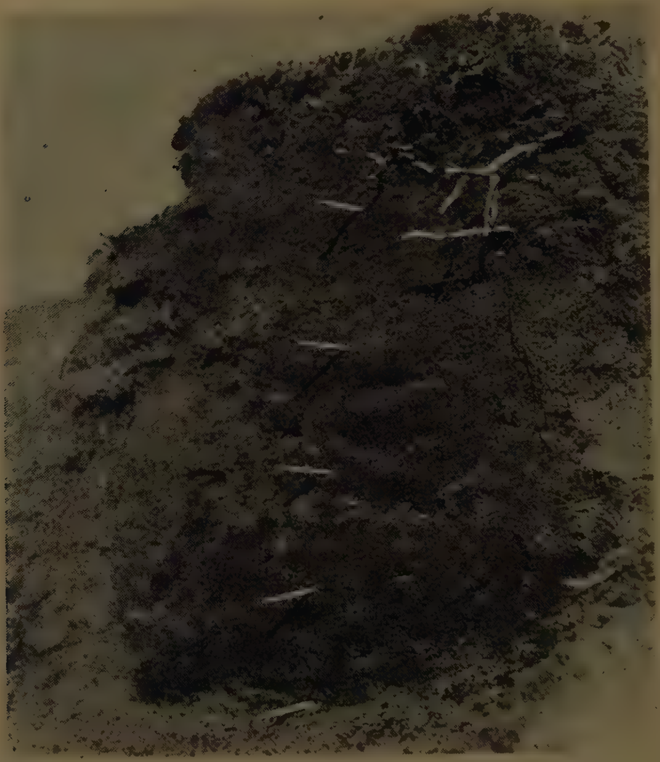

Fig. 8. Fra jordlaget på Midtre kam der dette er tjukkest (ca. 35 cm).

på Gråfjell og Sandåfjell et unntak idet den viser ei isrørsle som var mer uavhengig av relieffet. Isen må naturligvis her ha gått opp til minst 1550 m o.h. (Vi rekner da, både nå og seinere, med den høgda som landmassene har i høve til havet i dag.) Men da terrenget nord for dette fjellpartiet er lågere, møtte ismassene under rørsla nordover liten motstand. Det er derfor ikke sikkert at ismassene her gikk noe vesentlig høgre enn til 1550 m o.h.

De ismassene som trengte fram mot Trollheimen må stort sett ha hatt rørsle mot nordvest. Dette kan en f.eks. slutte av den ting at på det geologiske kartblad Folldal (1936) går skuringsstripene i det vesentlige i den lei. Dette faller jo også godt inn i det bilde vi har av isrørsla sør og søraust for Dovre. Men ismassene har trulig ikke vært tjukkere enn at Trollheimen har vært en alvorlig hindring å passere. Isen er blitt pressa oppover Gjevilvassdalen og Storlidalen og fra Gjøra har den følgt Sunndalen i nordvestlig lei. Fra disse hovedtransportrennene har den greina seg utover i Trollheimen der det var lettest å komme fram. En isstraum må vi således sikkert ha hatt mellom Gjevilvasskammene og på begge sider av Blåhø, videre nedover Slettådalen og Fossådalen til Folldalen som så har overtatt transporten. Fra Storlidalen må isen ha trengt fram over To-vatna. Her har den møttes med en straum oppover Otdalen. Fra det sentrale Trollheimen har trulig samstundes en isbre trengt seg mot vest gjennom Nøstådalen. Som vi ser, vil alle påviste skuringsmerker få si naturlige forklaring ved å gå ut fra ei slik isrørsle. Dette kommer også til å gjelde skuringsmerkene ved Gråfjell og Sandåfjell. Både Otdalen og det parti av Storlidalen som ligger rett aust for dette fjellpartiet peker omtrent i nordlig retning. Det er da ikke så urimelig å tenke seg at også det mellomliggende isparti ble dirigert nordover.

De funn som er gjort av flyttblokker og deres høgd i terrenget gir ei god støtte for det inntrykk av isrørsla en får ved å studere skuringsmerkene. En isstraum fra Gjevilvatnet forbi Gjevilvasskammene og nedover Folldalen måtte avta i høgd over havet i rørsleretninga. Det samme måtte være tilfelle med breen oppover Storlidalen, breen som følgte Sunndalen nordvestover fra Gjøra og dens forgreining oppover Otdalen. Er det slik som vi har antatt ovafor, at isen ikke var mektigere enn at den stort sett søkte seg fram langs daler og skar, må vi vente å finne flyttblokker heilt opp til det nivå som svarer til den øvre isgrensa. For det er klart at det støtt måtte rase ned stein på isoverflata fra de ovaforliggende fjellsidene. Ved å studere den øvre blokkgrensa i dag, skulle vi altså kunne finne fram til den gamle isoverflata. Hvis vi gjør dette, kan jeg ikke se at det kommer fram noen urimeligheter. Så vidt jeg kan bedømme det, får vi fram gradienter i det øvre blokknivå som svarer omtrent til det en måtte vente. Rekner vi således fallgradienten i nord-sørlig retning fra Gråfjell—Sandåfjell til ei linje på høgd med Trollhetta, får vi — dersom vi antar at ishøgdene ved de to yttergrensene var 1550 og 1300 m o.h. og avstanden 24 km — ca. 1/100. For ei austvestlig midtlinje gjennom Gjevilvasskammene får vi under disse forutsetninger ei ishøgd på 1425 m o.h., noe som svarer svært godt til den observerte blokkhøgda. I Rondane var etter STRØM (1945)[8] og SUND (1945)[9] den maksimale ishøgda ikke større enn at isen så vidt rakk opp over toppene (knapt 2200 m o.h.). De ismassene som strømte fram over Trollheimen, kom vel stort sett fra søraust. Mellom Rondeslottet og Trollhetta er 110 km i luftlinje. Rekner vi med jamt fall på isoverflata fra 2200 m o.h. ved Rondeslottet til 1300 m ved Trollhetta, får vi en gradient på 1/122. Da vi jo måtte vente at isoverflata i de mer sentrale nedisingsområder hadde ei mindre helling enn i de perifere strøk, må gjennomsnittsgradienter fra Rondeslottet til Trollhetta ha vært mindre enn den vi hadde innen Trollheimenområdet. Dersom derfor en gradient på 1/100 er så noenlunde rett her, må også gradienten 1/122 for den lengere distansen være det. Vi har riktignok rekna gradienten i Trollheimen mot nord og den andre mot nordvest. For samme område vil vel en gradient i nordvestlig retning bli litt større enn en i nordlig, men ikke så mye at det vil ha noe å si for vårt resonnement. Med en gradient noe større enn 1/100 i nordvestlig retning fra Gråfjell—Sandåfjell til ei linje på høgd med Otthø vil vi få et fall på isoverflata på vel 100 m. (Gradienten 1/100 vil gi omtrent akkurat 100 m.) Dette stemmer jo også bra med at det ikke ligger flyttblokker over ei høgd på ca. 1400 m o.h. i traktene vest for Otthø.

PER HOLMSEN har naturligvis rett i at vi neppe kunne vente å finne flyttblokker på de høgste toppene i Trollheimen om landskapet hadde vært totalnedist under siste istid. Men jeg tror nok at vi ville ha funnet dem høgre opp i terrenget enn vi gjør i dag. Et enkelt resonnement vil sannsynliggjøre dette: Om isen hadde vært mektig nok til å trenge seg fram noenlunde uavhengig av relieffet — noe som vel måtte ha vært tilfelle om den strakte seg opp over de høgste toppene — ville sikkert rørsla over det heile ha vært stort sett mot nordvest. Under denne forutsetning ville f.eks. ismassene som passerte Svarthetta (1572 m o.h.), gå videre mot Trollhetta. Det ville da ha vært rimelig at en og annen flyttblokk ble med fra Svarthettaområdet og ble lagt igjen ved Trollhetta når firngrensa under avsmeltinga hevde seg slik at isen smelta fra toppene. Vi skulle da ha venta å finne blokker høgre enn 1300 m o.h. Og disse blokkene ville ikke ha vært forsvunnet i dag som følge av forvitring. Dette går tydelig fram av det faktum at de blokkene som ligger høgst i terrenget nå, er svært lite påvirka av forvitringa.

Er mi tolking av tilhøva rett, er det klart at vi har hatt mange nunatakkområder i Trollheimen under siste istid. I det området vi her særlig har sett på, blir det f.eks. tilfelle med Trollhetta, Snota, Svarthetta, Storlifjellet og Gjevilvasskammene. Og i det området som HANS HOLTEDAHL omtaler fra Gjevilvatnet og sør- og sørvestover til Sunndalen, stakk Hornet, Okla, Gråfjell, Sandåfjell, Kråkvasstind, Svartdalskollen, Nonshø, Otthø, Sommerungsnebba, Klingfjellet og det navnlause fjellpartiet nordvest for Klingråket opp over den sammenhengende isoverflata.

Nå er det ikke slik å forstå at vi har å gjøre med absolutt isfri nunatakkområder. Vi må huske på at den klimatologiske snøgrensa sikkert stort sett lå temmelig lågt under maksimum av siste istid. Lokalglasiasjon måtte derfor komme til å gjøre seg sterkt gjeldende. Breer og fonner av stedlig karakter lå over vesentlige områder i direkte kontakt med «storisen», dvs. ismassene som fra isdeleren og de sentrale nedisingsområder i søraust trengte fram gjennom Trollheimen. Egentlig isfritt kunne det bare bli i områder der snøen blåste av, der det var for bratt til at den kunne få feste, eller muligens områder med særlig gunstig sørvendt beliggenhet der lokalt den klimatologiske snøgrensa kunne tenkes å ligge særlig høgt, slik at den kom over vedkommende lokalitet. Den siste typen isfritt område vil på en måte være en lokal parallell til EILIFS DAHLS nunatakk av skandinavisk kystfjelltype.

Det kan ikke være å vente — som HANS HOLTEDAHL meiner — at vi i Trollheimen skulle kunne finne en markert topografisk grense mellom et område som var dekt av «storisen» og eventuelle nunatakker. Da den klimatologiske snøgrensa — i alle fall i det vesentlige — lå under overflata av ismassene, bøyde isen seg konkavt oppover langs fjellsidene. Vi fikk derfor ingen konsis grenselinje oppad for isen. Og da «storisen» dessuten over store områder gikk direkte over i lokalglasiasjonen, ville det ikke være noen vesentlig skilnad på de fysisk-kjemiske tilhøva for forvitring over og under grensa for «storisen». Ei slik topografisk grenselinje kan en neppe få uten at kontaktlinja mellom is og fjell ligger under den klimatologiske snøgrensa, noe som bare i ubetydelig grad kan ha vært tilfelle i Trollheimen under maksimum av siste istid.

De områda som HANS HOLTEDAHL har undersøkt mellom Gjevilvatnet og Sunndalen, har jeg bare overfladisk kjennskap til. Men ved å se på HOLTEDAHLS bilder og å studere kartet, får en absolutt inntrykk av at topografien nettopp er slik som en måtte vente det dersom vesentlige deler av landskapet lå over «storisen» og ble utsatt for virkninga av lokalglasiasjon. De skarpe formene og den sterke forvitring i de høgre nivåer får da sin naturlige forklaring. Også her må da naturligvis bare høgst avgrensa områder ha vært isfri i egentlig forstand. I heile landskapet over «storisen» har de tærende krefter fått godt tak. Lokale breer av forskjellig type kunne arbeide under heile istida inntil den klimatiske snøgrensa ble hevd over næreområda. Under snø og fonner kunne vi få en intens frostforvitring, og da disse siste agenser har en forholdsvis liten eroderende effekt, ville det seinere smelte fram betydelige mengder lausmateriale under dem. Endelig ville også de heilt isfri partia gjennom lang tid være sterkt utsatt for de tærende krefter så vi fikk danna urer og blokkhav.

Når det gjelder jordlaga på Gjevilvasskammene, trur jeg at vi må kunne slå fast at de i det vesentlige vitner om ei mektig forvitring *in situ*. Dersom det er rett at siste istids iskappe hadde ei høgd på vel 1425 m o.h. i Gjevilvasskammene, lå jo platået på Midtre kam der jordlaget ligger, atskillig over den sammenhengende isoveroverflata. Men her var neppe snøfritt. SØRENSEN har nok rett i at det inntil for kort tid sia lå ei snøfonn over platået. Under istida lå den nok sikkert der. I den postglasiale varmetida må vi rekne med at den var borte, for så å komme igjen i subatlantisk tid. Hevinga av snøgrensa i våre dager har fått fonna til å forsvinne på ny. Da slike fonner har liten evne til erosjon, samtidig som det foregår atskillig frostsprengning under dem, har trulig snøfonna på toppplatået av Midtre kam både virka konserverende på et forvitringsdekke som måtte ha eksistert på forhånd og rimeligvis også utvikla det videre. Jordlaga på Hemre kam synes å ha befunnet seg så å si i overflata av siste istids isdekke, slik at isen var så tynn at den ikke hadde noen eroderende effekt. Det eventuelle tynne isdekket på kammen kom til å ligge i ro mens ismassene bevegde seg nordvestover på begge sider av fjellryggen.

Det er imidlertid vanskelig på grunnlag av jordlaga på Gjevilvasskammene å dra noen sjølstendige slutninger. Det er et faktum at de er knytta til meget lause bergarter, noe som går tydelig fram av bildet fra Hemre kam (fig. 9). Vi ser hvordan erosjonen har gjort kammen salforma nettopp der de lauseste laga er. Det er de samme

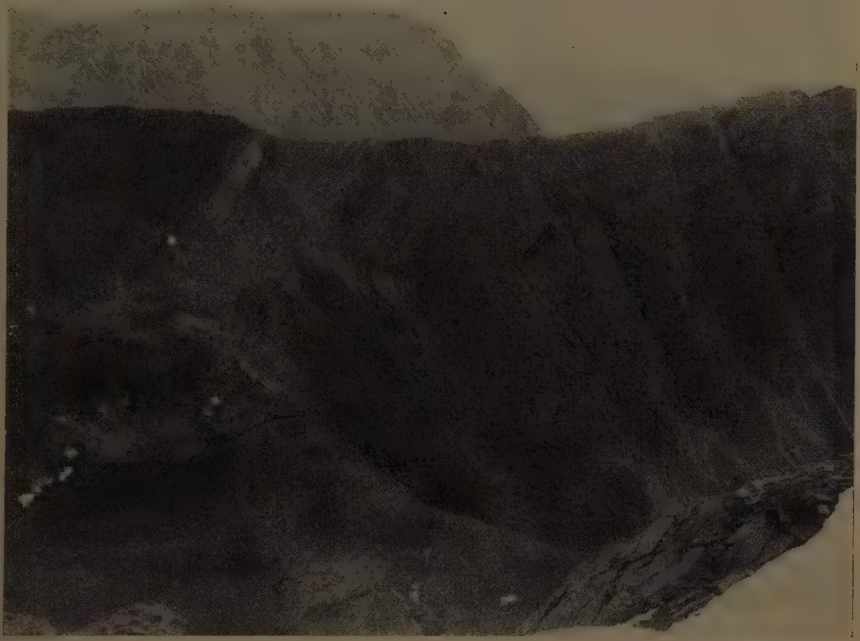

Fig. 9. De sterkt omtalte jordlaga på Hemre og Midtre Gjevilvasskam er knytta til lause skiferbergarter. På dette bildet ser en det tydelig for Hemre kam sitt vedkommende. Laga ligger oppå den salforma delen av ryggen, der bergartene er særlig lause, så forvitring og erosjon har fått gjort seg sterkt gjeldende.

lause laga som fortsetter over i Midtre kam. En kjenner jo også ellers alt for lite til hvor fort forvitringa av slike bergarter kan foregå under forskjellige tilhøve.

Det materiale som jeg har lagt fram ovafor, kan kanskje ikke på noe punkt sies å være direkte prov for at vi har hatt nunatakkområder i Trollheimen under siste istid. Men en må i alle fall ha lov til å si at de geologiske tilhøva gir sterke indisier på at isen ikke gikk over de høge toppene.

De botaniske undersøkelser som SØRENSEN og særlig GJÆREVOLL har foretatt i de seinere år, synes absolutt å gi støtte for den oppfatning at det har vært nunatakkområder i Trollheimen under siste istid. Den sterke konsentrasjon av «glasiale overvintrere» ved og i den umiddelbare nærhet av Gjevilvasskammene, vil få sin naturlige forklaring ved å anta at kammene (og Blåhø) hadde nunatakkområder. Dersom vi i det heile tatt har hatt nunatakkområder der hardføre planter kunne «overvintre», må nettopp Gjevilvasskammene og Blåhø ha vært særlig gunstige representanter. De har høgtliggende bratte, sørvendte skråninger over det nivå som en må anta «storisen» nådde opp til, og der er ypperlig jord som egner seg nettopp for slike plantetyper. Sjøl om vi må rekne med at den klimatologiske snøgrensa i de omliggende strøk lå lågere enn de bratte skråningene der plantene eventuelt må ha levd, ville naturligvis ikke de bratte avhell gi feste for snøen. Slike sørvendte bratte skråninger vil sikkert lokalt heve den klimatologiske snøgrensa ganske betydelig, og en kan naturligvis på slike steder ha temperaturtilhøve om sommeren som en ellers normalt finner i mye lågere nivåer.

Skal jeg til slutt kort resymere de slutninger jeg mener vi har rett til å gjøre på grunnlag av det som er framholdt i denne artikkelen, må det bli at det er overveiende sannsynlig at vi hadde nunatakkområder i Trollheimen under heile siste istid. Over «storisen» hadde vi lokalglasiasjon. Men i visse områder der vinden blåste snøen vekk, eller der skråningene ble for bratte til at snøen fikk feste, var det heilt isfritt. I bratte sørvendte skråninger kunne sommertemperaturen sjøl midt under siste istid ligge så høgt at hardføre planter kunne leve på laglige steder.

ENGLISH SUMMARY.

In an article in "Naturen" (nr. 3 1949) N. A. SØRENSEN reports the presence of some layers of stoneless soil on two mountains in Trollheimen — Hemre Gjevilvasskam and Midtre Gjevilvasskam — at heights of respectively 1400 and 1640 m above sea level; this he supposes to be from the last interglacial period. He therefore believes that the last ice sheet did not reach higher than 1350 m on these two mountains. SØRENSEN also draws attention to various botanical data which in his opinion support this theory. In this connection he stresses particularly his finding of *Arenaria norvegica* on Hemre Gjevilvasskam.

SØRENSEN has since continued his investigations in Trollheimen and has extended them westwards to the sea. These investigations are meant to support the theory of nunataks. The results are not yet published.

The geologists HANS HOLTEDAHL (1949, 1950)[3] and PER HOLMSEN (1951)[4] have participated in the discussion started by SØRENSEN. In HOLTEDAHL'S opinion Trollheimen must have been totally glaciated during the last glacial period. HOLMSEN is more non-committal and is unwilling to draw any definite conclusion until an objective method has been found for fixing the age of the material of disintegration.

OLAV GJÆREVOLL has worked on the flora of Trollheimen. He believes that several of the plants which he and other investigators have discovered in Trollheimen (fig. 1) must have survived there during the last glacial period and that Hemre- and Midtre Gjevilvasskam were the "wintering place".

Own Investigations.

The present author has since 1946 worked on various problems of quaternary geology in Trollheimen, though unfortunately less systematically than would have been desirable. Some of the results are reported in this paper. Erratic blocks were not found above 1300 m on Trollhetta and Snota. On Hemre Gjevilvasskam the highest erratic block found lay about 1420 m a.s.l. The ice-scourings mentioned were all found below 1200 m. The layers of soil on Hemre- and Midtre Gjevilvasskam (figs. 5, 6, 7, 8) have been more closely investigated; they proved to consist of soil of disintegration formed *in situ*.

Discussion.

A study of the relevant data collected by myself and others, definitely leads me to believe that the movements of the ice-masses from the last ice age in Trollheimen were determined by the relief: the ice has been able to force its way only along valleys, passes and glens. The ice-scourings on Gråfjell and Sandåfjell (1550 m a.s.l.) are an exception, manifesting an ice-movement which is more independent of the relief. As, however, the territory north of this part of the mountains is lower, the ice-masses met with little resistance when moving northwards. It is therefore not certain that the ice-masses reached much higher than 1550 m a.s.l. in this area.

If it were the case that the ice-masses had passed over Trollhetta, it is likely that erratic blocks would have been found, if not actually on top, then at any rate higher than we find them to-day. If the ice had been powerful enough to force its way fairly independently of the relief the movement would certainly have been largely northwest. Ice-masses which for instance passed Svarthetta would have left erratic blocks higher than those noted at 1300 m on Trollhetta.

When we study the gradient which the surface of the ice must have had if it was to carry and deposit erratic blocks at its highest extreme limit, we find reasonable values — about 1/100. In Rondane, according to STRØM (1945)[8] and SUND (1945),[9] the ice only just covered the tops (2200 m) at its maximum height. If we reckon with a fall in the surface of the ice from 2200 m near Rondeslottet to 1300 m near Trollhetta we get an average gradient of 1/122, a value which is not improbable.

In the author's opinion there is nothing to contradict the assumption that a series of nunataks were to be found in Trollheimen during the last glacial period. This does not mean that there must have been nunatak areas completely free from ice: the climatological snow-limit during the last glacial period may have been fairly low, resulting in a very strong local glaciation above the level of the inland ice. The only areas which could be really free from ice would be those where the snow was blown away, or those which were too steep for the snow to collect or, possibly, those which benefited from a particularly southerly situation where locally the climatological snow-limit might be expected to be especially high.

It is not to be expected that in Trollheimen we should find a distinct topographical border-line between an area covered by the inland ice and possible nunataks, as the chemical and physical conditions for erosion above and below this border-line are essentially the same. Such a topographical border-line would only be likely to be found in areas where the border-line between ice and mountains lies below the climatological snow limit, which could to only a limited degree have been the case in Trollheimen during the last glacial period.

The areas investigated by HANS HOLTEDAHL between Gjevilvatnet and Sunndalen I know only superficially, but I am inclined to believe that the topography of the terrain is just such as might have been expected if large parts of the ground lay above the inland ice and were exposed to local glaciation.

Merely on the basis of the presence of the soil layers on Hemre- and Midtre Gjevilvasskam it is difficult to draw any conclusion regarding the thickness of the last ice sheet, since these layers are found on very loose types of rocks.

The botanical investigations recently carried out by SØRENSEN and even more those carried out by GJÆREVOLL support the theory of nunataks in Trollheimen during the last glacial period. On steep southern mountain-sides summer temperatures could probably even with the ice cover at its maximum thickness be sufficiently high as to make possible the existence of hardy plants in favourable spots.

TEXT-FIGURES.

Fig. 1. Part of the Trollhetta Quadrangle (42 C). Numbers indicate locations where some of the most important «glacial survivors» were found. 1) *Papaver radicatum*, 2) *Arenaria norvegica*, 3) *Luzula arctica*, 4) *Sagina caespitosa*, 5) *Taraxacum dovrense*, 6) *Nigritella nigra*,

Fig. 2. The highest situated erratics found in Trollhetta at 1300 meters a.s.l. Weathering slight.

Fig. 3. The highest situated erratic found in Hemre Gjevilvasskam, at 1421 meters a.s.l.

Fig. 4. Glacial shatter-marks seen at 1184 meters a.s.l. on Trollhetta. The marks indicate ice-movement towards the northwest, along Folldalen.

Fig. 5. The soil layer on Hemre Gjevilvasskam.

Fig. 6. Section of the edge of the uppermost soil-layer on Hemre Gjevilvasskam. No Marked boundary visible in the section. Assumed to have been formed by weathering.

Fig. 7. A small trench dug near the soil-layer on Midtre Gjevilvasskam. This also is a weathering-product, with no marked boundary to underlying rock.

Fig. 8. The soil-layer on Midtre Gjevilvasskam at its thickest point (ca. 35 cm).

Fig. 9. The much-discussed soil-layer on Hemre- and Midtre Gjevilvasskam is connected with the underlying schistous rocks — clearly seen in this picture in the case of Hemre Gjevilvasskam. The layer lies above the saddle-shaped part of the ridge, where the rocks are so loose that weathering and erosion have been especially active.

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A PETROGRAPHICAL AND MINERALOGICAL STUDY OF TWO HIGH ALTITUDE SOILS FROM TROLLHEIMEN, NORWAY

BY

HANS HOLTEDAHL.

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Abstract. Sedimentary petrographical methods are used in the study of two soils on the summit-plateau of Midtre Gjevilvasskamm, Trollheimen, 1640 meters above sea level. The soil-layer of the plateau, i.e. section I, is about 30 cm thick and overlies a hornblendeschist. It is a well-sorted sand, with medians of 0.17 mm and 0.19 mm. The silt and clay content is very small, showing a slight increase in the lower part of the section, probably as a result of mechanical illuviation. From the mineral studies and the mechanical composition it is assumed that the soil is a weathering product of the underlying rock. The soil is thought to be mainly a product of physical weathering, but a small amount of chemical weathering is presumed to have taken place.

The soil of section II is also from the plateau, north of section I. Under a veneer of stones there is a well-sorted sand which with increasing depth becomes gradually more unsorted. The medians of the three soil samples taken from the top down to a depth of 70 cm, are 0.19 mm, 0.17 mm and 0.7 mm respectively. The silt and clay content of this section is much higher than that of section I, reaching a maximum at a depth of 25—30 cm. The latter is believed to be due to mechanical illuviation. From the study of the mineral- and rock material of the soil, both macroscopically and microscopically, and of the mechanical composition, the soil is believed to be of glacial origin, but highly disintegrated by weathering in its upper part. As in section I the weathering is believed to be mainly physical.

The age of the soils is not thought to be older than the last (Wisconsin) glaciation.

Introduction.

In papers dealing with the geology of the Opdal—Trollheimen area in south-western Norway (H. Holtedahl 1949 a, 1949 b, 1950) the writer pointed out the great rock disintegration found on summits and summit plateaux and on cirque walls and areas recently uncovered from shrinking snowfields. It was also assumed that the rock disintegration had taken place after the ground was freed from the last ice sheet, as striae and erratics were occasionally present. Large areas were found to be covered with rock debris of the native rock, and the intensity of disintegration depended greatly upon the structure and the type of rock, so that the greatest disintegration had taken place in areas where the rocks were schistous flagstones, quartzites, biotite-gneisses, hornblende-schists and mica-schists. The weathering was presumed to be mainly physical, with frost action as the main agent. The effect of frost action on rocks of various types was found to be very different. Flagstones and quartzites split along their mica-bearing schistosity planes, which were usually more than 5 cm apart, and the result was a mass of slabs or flags, sharp-edged. When the rocks consisted of hornblende-schists, mica-schists and biotite-gneisses the frost weathering resulted in all stages of disintegration from blocks to sand and silt, with the finer material predominant.

On mountain plateaux where the ground was fairly level, and movement of the debris by solifluction or running water was small, residual soil of some thickness was able to form. The soil was often inhabited by mosses, lichens, and a few higher plants.

The presence of residual soils or highly disintegrated rocks, and the absence of glacial striae, roches moutonnées and erratics, has often been used as an indication that the areas were not overridden by the ice sheet during the last glaciation. KALDHOL 1932, NORDHAGEN 1933, 1935; UNDÅS 1938, DAHL 1946, 1947, and others). In Norway, it has been claimed, such unglaciated areas have existed both along the coast, and further inland on high ground, as nunataks. Despite the fact that botanical data may point to the existence of unglaciated areas, the geological data are not conclusive, and the problem is far from finally resolved.

It has recently been argued that certain mountain-peaks in Trollheimen have escaped the glaciation of the last ice sheet, and that these peaks existed as nunataks. (N. A. SØRENSEN 1949, A. GRØNLIE 1950). The lack of erratics and striae, and the presence of highly weathered rock on the summit plateaux of Hemre- and Midtre Gjevilvasskam was supposed to be an indication of this. The hypothesis was, however, found to conflict with observations made by the author in areas investigated further west, whose highest points show definite signs of glaciation by the inland ice. (H. HOLTEDAHL 1950). Other work in the area has been done by P. HOLMSEN (1951), who studied the soil accumulations described by Sørensen and Grønlie, and came to the conclusion that the detritus was a weathering product deposited by wind action.

The author felt that a more detailed analysis of the soil accumulations would be of interest, both from the point of view of the quaternary geology and from that of weathering and soil formation in an alpine region. Owing to the inaccessibility and the lack of importance for cultivation purposes, very little is known as to the conditions of soil formation at high altitudes.

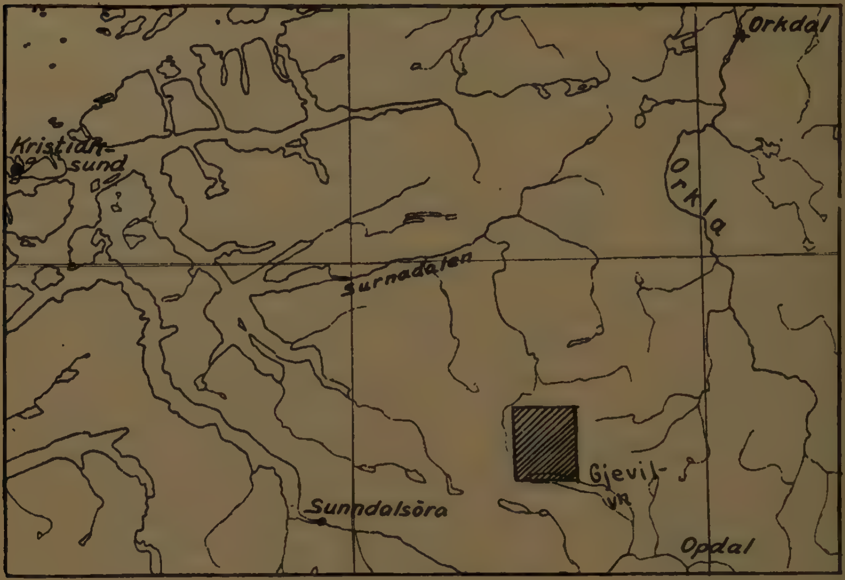


Fig. 1 a. Situation map.

The area of investigation.

The area investigated constitutes, as is shown on the map (fig. 1 b), the southern part of a summit plateau to the north of Lake Gjevilvann and about 1640 m above sea level. The plateau, which is roughly one square kilometer in area and is probably a remnant of the old pre-quaternary land surface, is highly dissected by cirque erosion, with well developed cirques on both the southern and northern sides. Southwards the plateau extends into a very ragged arrête formed by sideward erosion of cirques. A nevé, the size of which has decreased greatly in recent times, is situated on the eastern slope of the plateau. The headwalls and sidewalls of the two cirques, south and south-east of the plateau, are steep, often precipitously so.

The surface of the plateau is to a great extent covered by debris from the native rocks; on the most southern part an extensive soil accumulation is found, with a sparse vegetation consisting of moss, lichen and some higher plants. The geology of the area has not as yet been investigated in any detail, but the rocks found are mostly

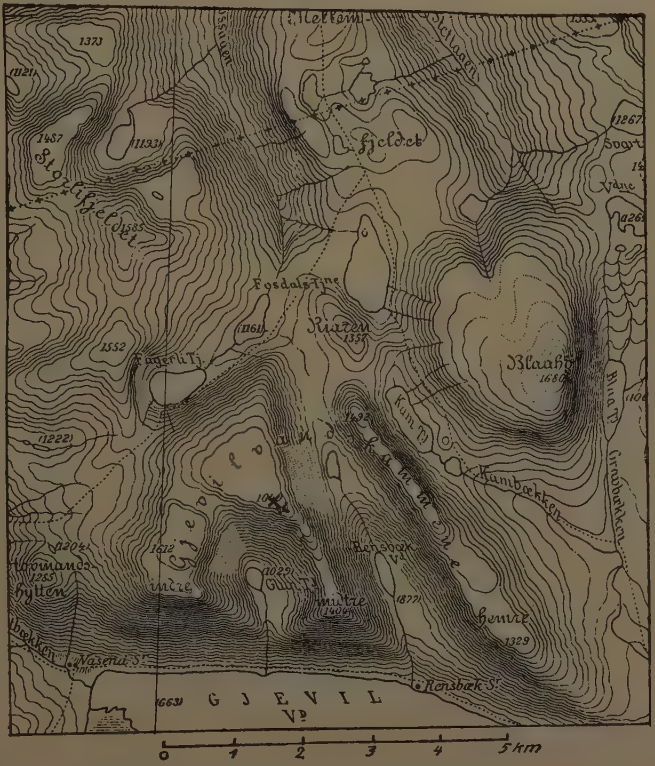


Fig. 1 b. Map showing area of investigation. The positions where samples were taken are indicated with crosses.

biotite-gneisses, hornblende-schists and mica-schists. The strike of the rocks is E.N.E. and the dip is about 30 degrees towards N.N.W.

At the southern tip of the plateau the wind has to some extent eroded and removed the soil, and here good sections can be studied. The soil-layer is about 30 cm thick, rusty brown in colour and with a sandy texture. The underlying rock is highly physically weathered, and there is no distinct boundary between soil and rock. There is no profile development visible. Samples were taken from the top layer, 0—5 cm below the surface, and from 5—25 cm, as well as from the underlying rock. This section is later referred to as section I.

To the north of this section the surface of the plateau is covered with cobbles and rather coarse debris. Here a ditch was dug down to a depth of about 80 cm below the surface. Under the top layer of coarse detritus the material is sandy, with an increasing amount of coarse material downwards. The colour of the material is a rusty brown, and there is no profile development to be seen. Samples were taken from 0—25 cm, 25—30 cm and 65—70 cm below the surface. The underlying rock was not reached. This section is later referred to as section II.

Climate of the region.

Knowledge of the present climate in the investigated area is important to an understanding of the present geological processes. Meteorological stations are, however sparse in this area, and the actual climatic conditions cannot be determined with any accuracy. The nearest stations where meteorological data are available are Lønset (503 meters above sealevel [a.s.l.]) about 10 km to the south, Opdal (605 meters a.s.l.) about 15 km to the south-east, and Inderdal (390 meters a.s.l.) about 30 km to the west. These stations do not, however, keep temperature records. The nearest stations where temperature records can be obtained are Sunndal (200 meters a.s.l.) about 20 km to the south-west, and Berkåk (451 meters a.s.l.) about 25 km to the north-east. A few observations from these stations are given below. (Geofysiske Publikasjoner, Vol. XIV, Det Norske Vid.Ak. Oslo, 1936—1943).

Mean temperatures 1861—1920 (in C.°)

Station	Meters a s.l.	Jan.	Feb.	Mars	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
Sunndal	200	-3.7	-3.2	-1.5	2.8	7.7	11.7	13.6	12.4	8.8	4.1	-0.9	-3.5	4.0
Berkåk	451	-5.6	-4.8	-3.3	1.4	5.9	10.4	12.5	11.0	7.6	2.3	-2.5	-5.4	2.5

If we take a fall in temperature of 0.5 deg.C. per 100 meters' rise in altitude, the temperature at 1640 meters a.s.l. will be:

Station	Meters a.s.l.	Jan.	Feb.	Mars	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Year
Sunndal	1640	-10.9	-10.4	-8.7	-4.4	0.5	4.5	6.4	5.2	1.6	-3.1	-8.1	-10.7	-3.2
Berkåk	1640	-11.6	-10.8	-9.3	-4.6	-0.1	4.5	6.6	5.1	1.7	-3.7	-8.5	-11.4	-3.5

The monthly mean temperatures in the area investigated probably lie somewhere between the values estimated for Sunndal and Berkåk at 1640 meters a.s.l. It will be seen that only the months of May, June, July, August and September have mean temperatures above zero.

Mean number of days with frost ($< 0° C$) at Sunndal Station 200 meters a.s.l.

Station	Meters a.s.l.	Jan.	Feb.	Mars	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Year
Sunndal	200	28.7	26.6	27.3	19.7	6.1	0.1	—	—	1.3	13.4	25.2	29.1	177

The mean number of days with frost at the summit of Midtre Gjevilvasskamm must be considerably higher than for Sunndal, only 200 meters a.s.l. The author had personal experience of temperatures below freezing point at the end of July 1951, and in the middle of August 1952.

Mean monthly and annual amount of precipitation for the normal period 1901—1930 in mm.

Station	Meters a.s.l.	Jan.	Feb.	Mars	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Year
Sunndal	195	93	71	52	33	24	42	52	65	58	62	82	66	700
Lønset	503	66	48	35	18	21	38	56	70	47	44	56	41	540
Opdal	605	48	38	22	17	23	45	69	72	48	34	35	33	484
Inderdal	390	157	98	111	88	86	126	105	144	168	135	133	188	1469
Berkåk	425	56	45	38	29	33	67	67	106	68	45	51	42	647
M.Gjevilvasskamm approx.	1640	75	50	25-50	25	25	50	50-75	75	50	50	50	50	575-625

Mean monthly and annual amounts of precipitation are seen to be very different at the different observation stations. Particulary notable are the high values at Inderdal. The values estimated for the area investigated have been taken from the precipitation map of southern Norway. (Nedbøren i Norge. 1895—1943. Utgitt av Det Norske Meteorologiske Institutt. Oslo 1949).

From the climatological data it will be seen that the climate of the area investigated is severe and a great deal of the precipitation which falls on the mountains will fall as snow. On the plateau itself, however, snow will not be able to accumulate to any great extent, owing to the strong winds, but it will accumulate on the lee sides, particularly in the cirques.

Outline of methods used in studying the samples.

Thin sections were made of both the rock underlying the soil in section I and of neighbouring rocks. Detailed mechanical analyses were made, to show the size distribution of the particles in the various soil horizons. The procedure used was the sieving and pipette method described by Krumbein and Pettijohn (1938). Before the mechanical analyses were performed, organic matter was removed by treatment with 7 per cent H_2O_2. Heavy liquid specific gravity separations were done with bromoform (sp.gr. 2.8) on the size groups 1—0.25 mm, 0.25—0.065 mm and 0.065—c. 0.01 mm. To facilitate the mineralogical analyses, the minerals were cleaned with 25 per cent tartaric acid and then mounted on slides. Quantitative counts of the minerals were than made.

The loss of ignition was obtained by igniting, in each sample, both material treated with H_2O_2 and non-treated material, information hereby being gained about the humus content.

The pH value was determined with a pH-meter (radiometer). (The pH determination was kindly done by Miss Å. Omvik, Ph. D., at the Botanical Laboratory University of Bergen).

Description of soil section I.

Sample 1 = 0—5 cm, sample 2 = 5—25 cm.

MECHANICAL COMPOSITION OF THE SOIL.

The size frequency distribution of the samples is shown in the histograms and cumulative curves (figs. 2 and 3). As will be seen there is very little difference in the mechanical composition of the two samples. The main part of the material consists of grains with

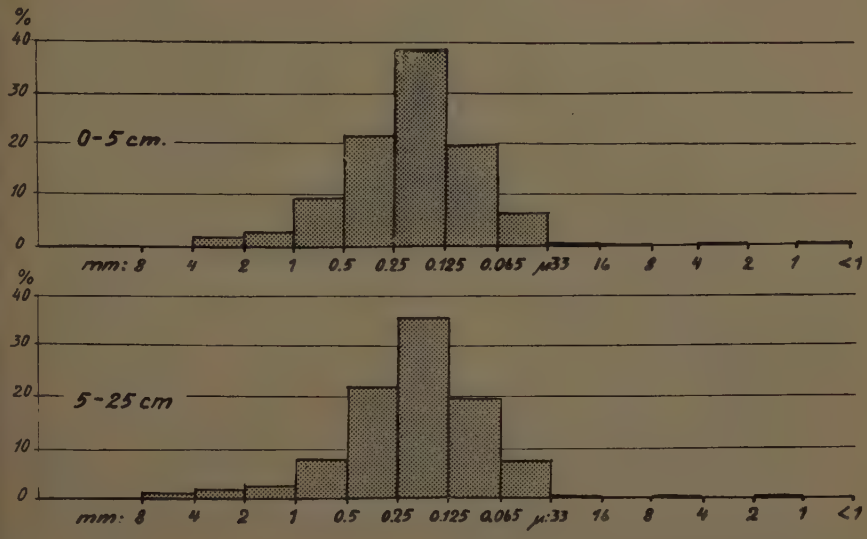


Fig. 2. Histograms of soil overlying rock at Midtre Gjevilvasskamm Section I, sample 1 and 2, 0—5 cm and 5—25 cm below the surface.

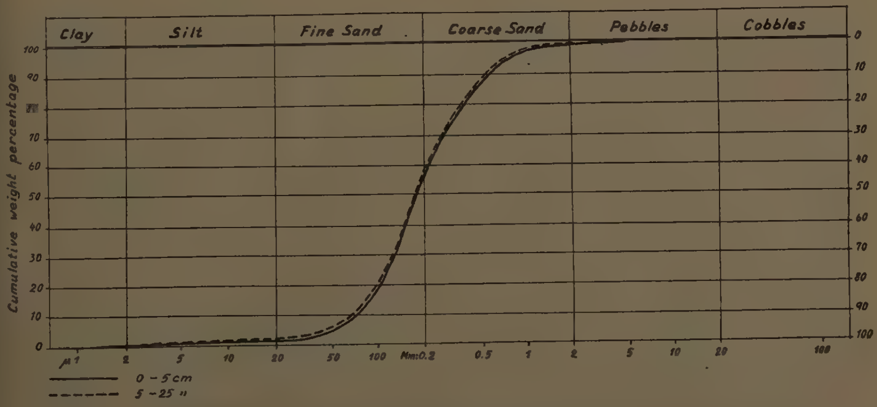


Fig. 3. Cumulative curves showing size composition of soil at Midtre Gjevilvasskamm. Section I.

diameters of between 2 mm and 0.02 mm, i.e. coarse and fine sand (sand and mo, Atterberg), with the maximum grade in the 0.125—0.25 mm range. The *median*, which is determinded by the point of crossing of the cumulative curve with the 50 per cent line, is 0.17 mm in the 0—5 cm sample, and 0.19 mm in the 5—25 cm sample (grovmo, Atterberg). The top layer contains no grains with a diameter above 4 mm; the underlying material has somewhat larger grains.

It will be seen from the histograms and the cumulative curves that the soil is well-sorted: the coefficient of sorting, So, which is the square root of the ratio of the quartiles, is for sample 1, 1.68 and for sample 2, 1.76.

In fig. 4 I the summation percentages of clay (< 0.002 mm), silt (lättler) (0.002—0.02 mm), very fine sand (finmo) (0.02—0.065 mm) and fine sand (grovmo) (0.065—0.125 mm) are shown as a function of depth. It will be seen that the clay and silt content of the soil, is very small in both samples, the clay content showing a very slight increase in the lower sample. The content of very fine sand and fine sand is also seen to be larger in sample 2.

MINERAL STUDIES.

Heavy liquid specific gravity separations were done with bromoform (s p.gr. 2.8) on the size groups 0.25—1 mm, 0.065—0.25 mm and 0.01—0.065 mm of each sample. A sample of the rock underlying the soil was crushed and heavy mineral separations performed as for the soil; the grade 0.25—0.5 mm was, however, used instead of 0.25—1 mm. The separations were done in filtering funnels, the stems of which were fitted with a short length of rubber tubing and pinch-cocks. Three separations of each sample were made, and the average was taken. The best separation was obtained in the grade 0.065—0.25 mm. The coarser grade contained a number of polymineral grains, and in the finer grade the particles were too fine to give a perfect separation by the method used. The percentage of heavy minerals was therefore best expressed in the size group 0.065—0.25 mm.

The results of the heavy mineral separations of the various samples are given in table 1.

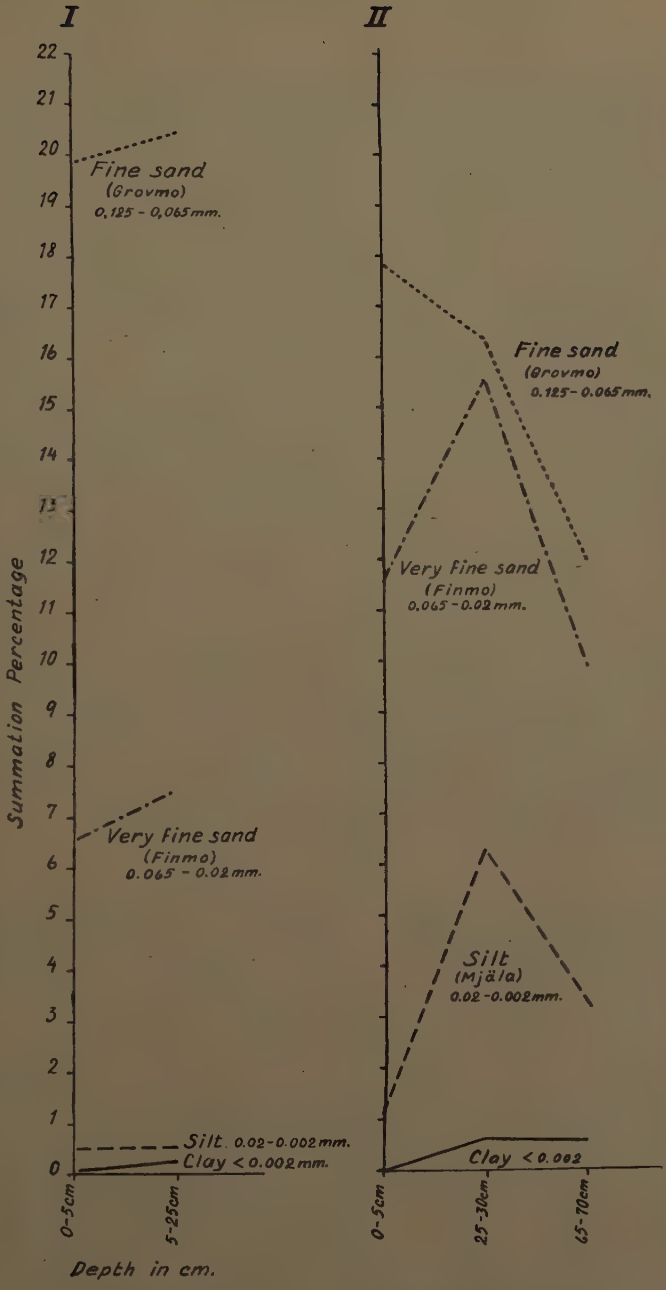


Fig. 4. Summation percentages of clay, silt, very fine sand, and fine sand shown as a function of depth.

TABLE 1.

Distribution of the heavy minerals (sp.gr.> 2.8) in the sand and silt fractions of the soil and underlying rock at Midtre Gjevilvasskamm, Section I.

Sample number	Depth in cm	Percentages of fractions		
		0.25—1 mm	0.065—0.25 mm	0.01—0.065 mm
1	0—5	52.2	55.2	62.4
2	5—25	52.0	57.3	64.5
3 (rock)	30	59.7 (0.5 mm)	61.4	78.8

Sample number	Depth in cm	Percentages of whole soil		
		0.25—1 mm	0.065—0.25 mm	0.01—0.065 mm
1	0—5	15.9	32.3	4.2
2	5—25	15.9	32.2	4.9
3	—	—	—	—

The separations of heavy and light minerals of the different fractions were then mounted on slides, and the different minerals counted. At least 500 grains were counted in each fraction, quartz and feldspar were counted together. The results of the examination of minerals are given in tables 2, 3, and 4.

TABLE 2.

Percentages of minerals identified in the coarse sand fraction (0.25—1 mm) of the soil and underlying crushed rock at Midtre Gjevilvasskamm, Section I.

Sample number	1	2	3 (rock)	1	2	3 (rock)
Depth in cm	0—5	5—25	30	0—5	5—25	30
Minerals:						
Hornblende	29.4	35.9	52.7	6	7	9
Biotite	21.7	28.5	0.3	6	6	1
Aggregates	22.0	11.5	14.1	6	4	4
Epidote and clinozoisite	1.6	—	—	1	—	—
Quartz and feldspar	25.4	24.1	32.9	6	6	7
	100.1	100.0	100.0			

For the sake of convenience the ranges in percentage have been replaced by single numbers in the right-hand side of the table.

1	0— 2 per cent	4	10—15 per cent	7	30— 40 per cent
2	2— 5 —	5	15—20 —	8	40— 50 —
3	5—10 —	6	20—30 —	9	50— 75 —
				I0	75—100 —

TABLE 3.

Percentages of minerals identified in the coarse- and fine sand fraction (0.25—0.065 mm) of the soil and underlying crushed rock at Midtre Gjevilvasskamm, Section I.

Sample number	1	2	3 (rock)	1	2	3 (rock)
Depth in cm	0—5	5—25	30	0—5	5—25	30
Minerals:						
Hornblende	53.5	50.3	59.7	9	9	9
Epidote and clinozoisite	2.4	8.3	2.0	2	3	1
Biotite	20.8	21.7	—	6	6	—
Muscovite	0.8	0.3	—	1	1	—
Titanite	0.2	—	0.3	1	—	1
Aggregates	1.2	—	0.3	1	—	1
Opaque	0.1	0.1	—	1	1	—
Quartz and feldspar	21.1	19.4	37.7	6	5	7
	100.1	100.1	100.0			

The mineral grains are mostly fresh. The hornblende is dark green, sometimes with a brownish stain. The biotite is a light golden variety, the colour probably being due to bleaching.

THE ROCK UNDERLYING THE SOIL.

The rock sample taken from underneath the soil layer is a dark hornblende-schist or schistous amphibolite, with porphyroblasts of quartz and feldspar. The rock is somewhat weathered, with rusty staining of the surfaces of the schistosity planes. The porphyroblasts are also to some extent rust-coloured.

In thin section the rock is seen to contain 50—60 per cent hornblende, 40—50 per cent quartz and oligoclase, and a few percentages of epidote and clinozoisite. A few grains of biotite are also present.

TABLE 4.

Percentages of minerals identified in the fine sand — silt fraction (0.01—0.065 mm) of the soil and underlying crushed rock at Midtre Gjevilvasskamm, Section I.

Sample number	1	2	3 (rock)	1	2	3 (rock)
Depth in cm	0—5	5—25	30	0—5	5—25	30
Minerals:						
Hornblende	54.7	38.2	68.0	9	7	9
Epidote and clinozoisite	4.3	24.7	10.1	2	6	4
Biotite	12.8	9.1	0.1	4	3	1
Muscovite	0.6	1.1	—	1	1	—
Chlorite	0.6	1.2	—	1	1	—
Titanite	0.6	—	—	1	—	—
Zirkon	1.1	0.4	—	1	1	—
Opaque	1.7	1.2	0.3	1	1	1
Quartz and feldspar	23.7	24.0	21.6	6	6	6
	100.1	99.9	100.1			

The rock has a marked parallel structure due to the good parallel orientation of the hornblende. The hornblende is a common hornblende, with pleochroism from light green to bluish-green and with $c : \gamma = 18°$. On the borders of the hornblende particles to other minerals, as well as on fractures, there is a marked brownish staining zone, due to oxidised iron. Quartz and feldspar show no sign of decomposition, occasionally their boundary lines have a faint staining zone like that of the hornblende minerals. The quartz and feldspar grains seldom exceed 0.6 mm in diameter. Epidote and clinozoisite are found commonly, either as large single individuals, or as clusters of very small particles. The grains have diameters of up to 0.3 mm, but are usually smaller.

THE COARSE GRADES OF THE SOIL.

In sample 1 the coarse grains are angular, and consist mainly of hornblende-schist as does the underlying rock. Some grains of feldspar and aggregates of quartz and feldspar are present. Micaceous quartz-feldspar schists are also seen and bleached biotite is found in flakes. Most of the grains are easily crushed between the fingers.

In sample 2 the grains are also angular and fresh and consist mainly of hornblende-schist. A few grains of a schistous mica-bearing quartz-feldspar rock are present. With a few exceptions the grains can all be crushed between the fingers.

IGNITION LOSS AND pH DETERMINATION.

The ignition loss of samples 1 and 2 was found by the following method: about 3 g. of the sample was mixed with 25 cc 7 per cent H_2O_2 and treated on a steam bath for 6—7 hours and then dried at 105° C. 3 g of the same sample was mixed with 25 cc distilled water and dried at 105° C. Both samples were weighed, and then ignited in an electrical furnace at c. 900° C. After igniting, the samples were weighed and the loss by ignition found. The result was:

Sample 1. Loss by ignition of untreated material.. = 4.35 per cent
Loss by ignition of H_2O_2 treated material = 2.10 —
Calculated humus content .... = 2.25 —
Sample 2. Loss by ignition of untreated material .. = 5.27 per cent
Loss by ignition of H_2O_2 treated material = 4.04 —
Calculated humus content .... = 1.23 —
Sample 3. The underlying rock was crushed and the rock powder ignited. The result was:
loss by ignition of crushed rock .... = 0.69 per cent

The pH value of the average soil was determined on a pH-meter and was found to be 5.3.

Description of soil section II.

Sample 1 = 0— 5 cm.
Sample 2 = 25—30 cm.
Sample 3 = 65—70 cm.

MECHANICAL COMPOSITION OF THE SOIL.

The size frequency distribution of the soil at various depths is shown in the histograms (fig. 5) and the cumulative curves (fig. 6). It will be noticed that the size distribution of the material varies greatly in the different soil horizons.

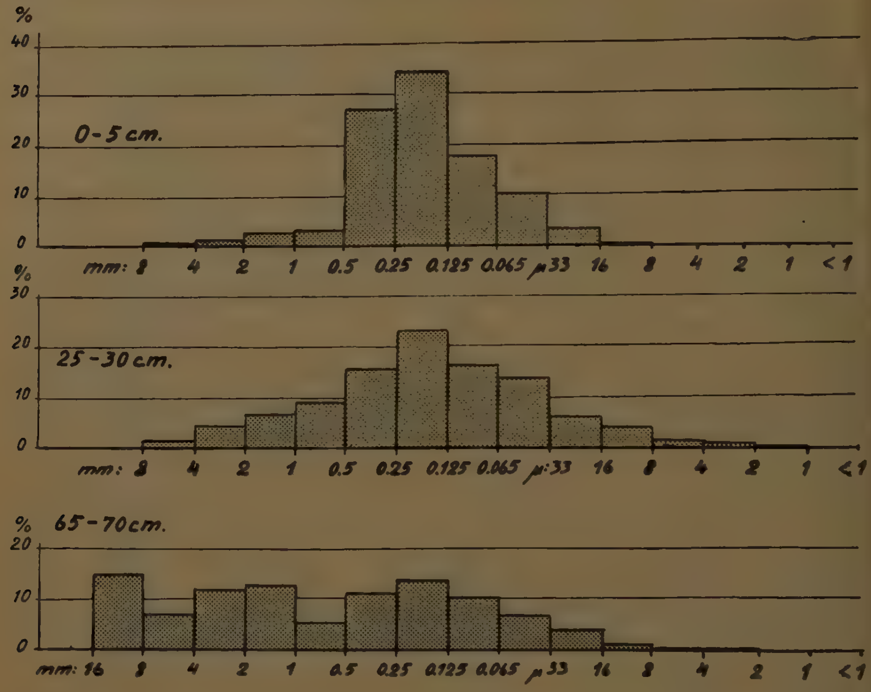


Fig. 5. Histograms of soil at Midtre Gjevilvasskamm. Section II, sample 1, 2, and 3, 0—5 cm, 25—30 cm, and 65—70 cm below surface.

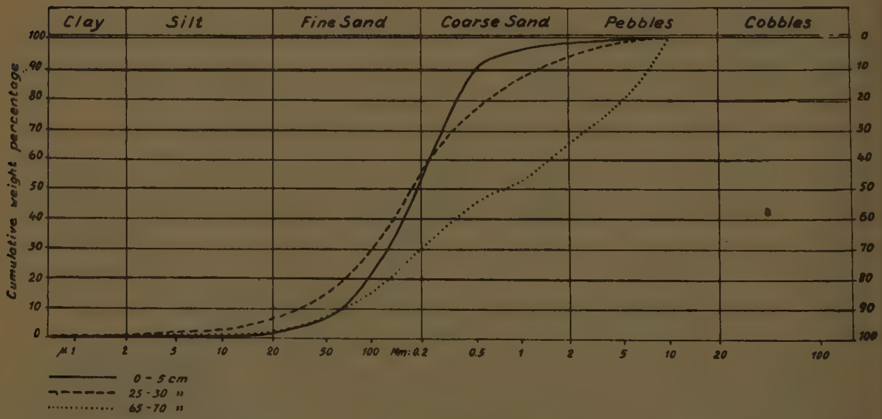


Fig. 6. Cumulative curves showing size composition of soil at Midtre Gjevilvasskamm. Section II.

The upper part of the section, i.e. sample 1, has material which is well-sorted. The sorting coefficient So = 1.73, which equals the So found in section I. The main part of the material consists of grains with diameters of between 0.02 mm and 2 mm, i.e. coarse and fine sand (sand and mo, Atterberg), with the maximum grade in 0.125—0.25 range. The median of the sample is 0.19 mm (grovmo, Atterberg).

The material in the middle part of the section, i.e. sample 2, is less well-sorted than that in the upper part, the So being 2.51. According to Trask (1932) a So value of less than 2.5 indicates a well-sorted sediment, 3.0 is normal, and a value greater than 4.5 indicates a poorly-sorted sediment. The main part of the sample consists of grains with diameters between 0.02 and 2 mm, with the maximum grade in the 0.125—0.25 mm range. The median of the sample is 0.17 (grovmo, Atterberg).

The lower part of the section, i.e. sample 3, consists of poorly-sorted material, the So being 4.76. The material consists to a great extent of pebbles and coarse and fine sand, and no maximum grade can be distinguished. The median of the sample is 0.7 mm (grovsand, Atterberg).

In fig. 4 II the summation percentages of clay (< 0.002 mm), silt (0.002—0.02 mm), very fine sand (0.02—0.065 mm) and fine sand (0.065—0.125 mm) are shown as a function of depth. Is is seen that the clay content is almost nil at the surface, but increases to 0.6 at a depth of 25—30 cm and shows a slight decrease at a depth of 65—70 cm.

In the silt and very fine sand grades there is a marked maximum at the depth of 25—30 cm. This is probably due to a downward movement and accumulation at this depth of particles belonging to the finer grades.

The percentage of clay and silt in section II will be seen to be considerably higher than in section I.

MINERAL STUDIES.

Heavy liquid specific gravity separations of the various sample were performed in the same way as for section I. Bromoform with a specific gravity of 2.8 was used, and the average values of four separations were taken. The result for the separations appears in table 5.

TABLE 5.

Distribution of the heavy minerals (sp. gr. > 2.8) in the sand and silt fractions of the soil at Midtre Gjevilvasskamm, Section II.

Sample number	Depth cm	Percentages of fractions		
		0.25—1 mm	0.065—0.25mm	0.01—0.065 mm
1	0—5	25.5	41.0	42.5
2	25—30	25.2	43.0	36.4
3	65—70	13.1	18.0	22.9

Sample number	Depth cm	Percentages of whole soil		
		0.25—1 mm	0.065—0.25 mm	0.01—0.065 mm
1	0—5	7.6	21.6	5.3
2	25—30	6.2	17.2	7.5
3	65—70	2.3	4.3	2.4

The separations of heavy and light minerals of the various fractions were mounted on slides, and the minerals counted. The results of the examinations are given in tables 6, 7, and 8.

TABLE 6.

Percentages of minerals identified in the coarse sand fraction (0.25—1 mm) of the soil at Midtre Gjevilvasskamm, Section II.

Sample number	1	2	3	1	2	3
Depth in cm	0.5	25—30	65—70	0.5	25—30	65—70
Minerals:						
Hornblende	18.8	0.5	0.1	5	1	1
Epidote and clinozoisite	5.4	19.1	1.5	3	5	1
Biotite	14.8	22.5	24.6	4	6	6
Muscovite	0.2	0.1	5.2	1	1	3
Aggregates	0.7	1.5	—	1	1	—
Opaque	—	1.8	1.2	—	1	1
Quartz and feldspar	60.2	54.6	67.5	9	9	9
	100.1	100.1	100.1			

To the right of the table the ranges in percentages have been replaced by single numbers, in accordance with the scheme used in section I.

TABLE 7.

Percentages of minerals identified in the coarse- and fine sand fraction (0.065—0.25 mm) of the soil at Midtre Gjevilvasskamm, Section II.

Sample number	1	2	3	1	2	3
Depth in cm	0.5	25—30	65—70	0.5	25—30	65—70
Minerals:						
Hornblende	23.0	4.4	0.5	6	2	1
Epidote and clinozoisite	16.4	33.8	4.7	5	7	2
Biotite	16.6	18.8	29.8	5	5	6
Muscovite	0.7	0.2	5.9	1	1	3
Rutile	0.1	—	—	1	—	—
Titanite	—	0.1	—	—	1	—
Aggregates	0.1	—	—	1	—	—
Opaque	0.2	1.0	0.2	1	1	1
Quartz and feldspar	42.9	41.7	58.9	8	8	9
	100.0	100.0	100.0			

TABLE 8.

Percentages of minerals identified in the fine sand-silt fraction (0.01—0.065 mm) of the soil at Midtre Gjevilvasskamm, Section II.

Sample number	1	2	3	1	2	3
Depth in cm	0—5	25—30	65—70	0—5	25—30	65—70
Minerals:						
Hornblende	25.0	4.6	0.9	6	6	1
Epidote and clinozoisite	23.3	21.5	24.4	6	6	6
Biotite	2.1	13.3	19.0	2	4	5
Muscovite	0.5	0.1	0.1	1	1	1
Rutile	0.1	—	—	1	—	—
Garnet	—	0.1	0.1	—	1	1
Zirkon	0.1	0.1	0.1	1	1	1
Opaque	1.6	0.6	3.6	1	1	1
Quartz and feldspar	47.5	59.7	51.8	8	9	9
	100.2	100,0	100.0			

THE COARSE GRADES OF THE SOIL.

In sample 1 six pebbles with diameters > 4 mm were analysed and found to represent the following rock types: 2 biotite gneisses, 1 schistose granite (granodiorite?), 1 massive suassurite gabbro with a large amount of epidote, 1 schistous muscovite quartz feldspar rock (flagstone), 1 hornblende-schist. Of coarse grains with diameters of between 4 mm and 1 mm, hornblende-schists and hornblende-bearing quartz-feldspar rocks are the most common, but some grains of muscovite-schist and quartz and feldspar are also present. Most of the grains are angular and can be crushed between fingers, but some of the larger pebbles have a slight sub-angularity and cannot be crushed with the fingers.

In sample 2 the material is more homogeneous than in sample 1 and consists mainly of flakes of a schistous quartz-feldspar rock containing a great amount of epidote, as well as biotite and hornblende. The grains are angular and can fairly easily be crumbled between the fingers.

Sample 3 consists mainly of biotite-bearing gneisses. Some pebbles of mica-chists, with a golden-coloured biotite, are also present. The pebbles are subangular-subrounded, with a somewhat polished surface, showing distinct signs of wear. The material cannot be crushed between the fingers.

IGNITION LOSS.

The ignition loss of two samples in section II was determined by the same method as was used in section I.

The result was:

Sample 2.

Loss by ignition of untreated material........ = 2.01 per cent
Loss by ignition of H_2O_2 treated material .... = 1.97 —
Calculated humus content .... = 0.04 —

Sample 3.

Loss by ignition of untreated material .... = 2.05 per cent
Loss by ignition of H_2O_2 treated material .... = 1.74 —
Calculated humus content .... = 0.31 —

Discussion.

The main objects of the present investigation are to throw light upon: a) the origin of the soils, and b) the weathering processes and the state of weathering of the soils.

SOIL SECTION I.

The origin of the soil.

As will be seen from the histograms and cumulative curves (figs. 2 and 3) the mechanical composition of the soil is very much the same in the two samples taken. The material is well-sorted, with sorting coefficients of 1.68 and 1.76, and with medians of 0.17 mm and 0.19 mm. It will also be seen that the clay and silt content is very low.

When we are seeking to determine the origin of the soil, deposition by wind action must be considered. According to Twenhofel wind deposists are comparatively well-sorted, the grains are usually well-rounded, minutely pitted and frosted. Coarse deposits are cross-laminated, and fine deposits are usually without lamination. (TWENHOFEL 1926 p. 65.) The perfection of sorting of course varies. Thus a dune sand near the coast, which originates from a water-sorted material, will show better sorting than a deposit which originates from unsorted material. The sorting will be dependent upon several factors, among which parent material and transport are the most important. A dunesand from the island of Vigra, west of Ålesund, W. Norway, was examined by the writer and showed very good sorting, with an So of 0.35, a median of 0.16 mm and the mode in the 0.125—0.25 mm class.

The material which is rolled on the ground by wind of a given current velocity is usually well-sorted, but owing to great variation in current velocity near the ground from place to place, the range in dimensions will vary considerably.

Udden (1894) found the range in dimensions of dune sand to be between 1/16 mm (0.063 mm) and 1 mm, with over 80 per cent of the material falling between 0.125 mm and 0.5 mm.

According to Sokolow (1894) there are no wind deposits known of which the grains exceed 4 or 5 mm in diameter, while the average dimensions are less than 1 mm. Pleistocene sand dunes near Sparta, Wisconsin, have dimensions ranging from about 0.2 mm to 0.6 mm, with a very small percentage above and below these limits (TWENHOFEL loc.cit.).

O. Holtedahl (1924) has examined Pleistocene dune sand at Romerike, north of Oslo. The range of dimensions of the grains is between 2.0 mm and 0.06 mm. No grains have diameters of over 2.0 mm, and only traces are found under 0.06 mm.

I. Högbom (1923) and G. Lundqvist (1940) have also examined wind deposits and find the modes mainly in the size ranges between 2 mm and 0.2 mm, and occasionally in the 0.2—0.06 mm class.

Pettijohn (1949 p. 44) summarizes data on wind-blown sediments supplied by Udden (1914), and finds that the modes are found most commonly in the 1/4—1/8 mm class (0.125—0.25 mm), but also commonly between 1/64 and 1/16 mm (0.016—0.065 mm).

To summarize these data: it will be seen that wind deposits have their range of dimensions mainly between 2 mm and 0.06 mm, and that grains with diameters of over 2 mm do not occur.

Turning now to the deposits of section I it will be seen that here there are particles present which have diameters of over 2 mm. In the upper part there are 1.4 per cent grains with diameters between 2 mm and 4 mm, and further down in the section there are 1.2 per cent grains with diameters of over 4 mm. Grains with diameters of up to 12 mm are found in this sample.

The co-presence of this coarse material with the finer, accordingly makes it appear highly unlikely that the soil could be eolian in origin. This assumption is also supported by a study of the minerals: this shows that the H.M. percentage is almost the same in the two soil samples, and does not vary much from that of the underlying rock. This homogenity of the soil is also very marked when the percentages of mineral species are estimated. This is especially well demonstrated in table 3: the percentage of hornblende, biotite and quartz-feldspar is very much the same in the two samples of the soil, while the content of epidote-zoisite is somewhat greater in the lower sample. It is also seen that the proportions of the different minerals in the three size fractions of the same sample show a certain relationship.

The lack of lamination, the freshness of the grains, which show no signs of wear, the lack of rounding and of pitted surfaces, and the lack of any distinct boundary to the underlying rock, are also facts which would seem to indicate that the material was not deposited by wind, though with short wind transport rounding and pitting may in some cases be insignificant. (O. Holtedahl 1924.)

A point probably in favour of a wind transport hypothesis is the presence of biotite in the soil, for the underlying rock was found to be poor in this mineral. This will be discussed below.

The conclusion having been reached that the soil is probably not a product of wind transport, the possibility of its being a product of weathering must now be considered.

The textural character of a weathered residium will vary with the nature of the parent rock as well as with the weathering process and the maturity of the end product. A weathering product which is due to advanced chemical weathering will consist of material which is mainly in the silt and clay grades. A weathering product of physical weathering is usually not so fine.

Pettijohn (1949 p. 375) gives some histograms showing the mechanical composition of some residual materials. The average of five samples of disintegrated granitic and gneissic boulders from glacial outwash shows good sorting: the range of grain sizes is from 8 mm to 0.033 mm, with the maximum grade in the 2—4 mm class. A sample of detritus from weathered gneiss in Columbia has grain sizes varying from 2 mm to 0.016 mm and with maximum grade in the 0.5—1 mm class.

S. Taber (1943 p. 1472) has examined disintegration products in Alaska, and gives examples of different stages of disintegration. Some of his analyses will be given below:

Sample	Location	>2 mm	2—0.84 mm	0.84—0.42 mm	0.42—0.25 mm	0.25—0.147 mm	0.147 0.074 mm	0.074 0.031 mm	< 0.031 mm
A	Candle Creek	7	12.1	20.7	17.0	23.2	16.2	2.0	1.8
B	Near Cushman Creek	9.6	16.6	20.0	9.4	6.1	5.6	20.8	11.9
C	Mt. McKinley National Park	11.3	5.7	7.1	6.9	8.3	39.7	6.1	14.9

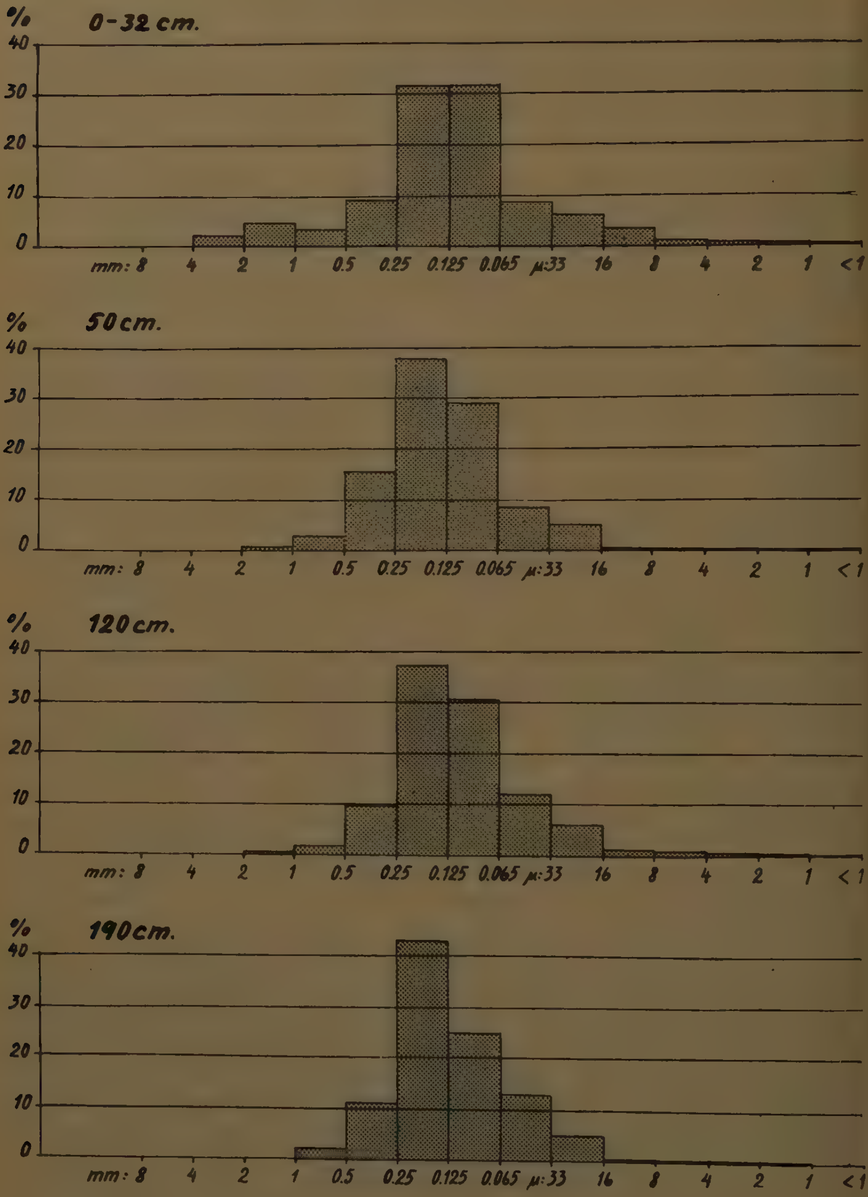


Fig. 7. Histograms of residual soil formed by weathering of mangerite-syenite, Holsenøy near Bergen.

Sample A represents the product of frost disintegration resulting from the exposure of slate for only 1 year. Sample B was obtained from a slope at an elevation of 1200 feet, and sample C from a slope at an elevation of about 3000 feet, both being from an unglaciated area and derived from schists.

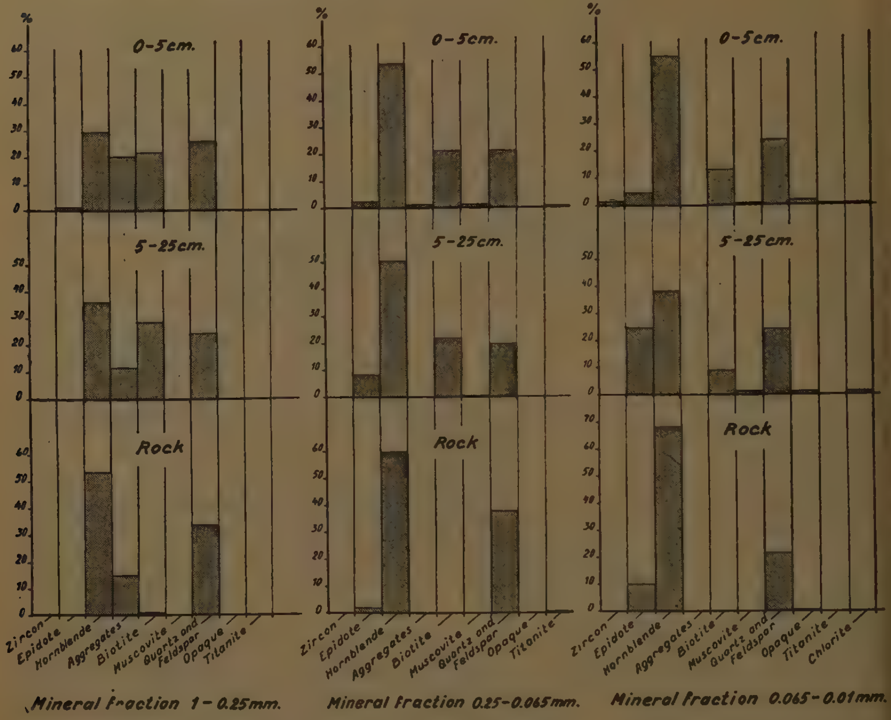
The writer examined a much disintegrated massive mangerite-syenite from Holsenøy, north of Bergen, This rock had disintegrated *in situ* to a depth of more than 2 meters, and since the area was glaciated during the last glaciation the disintegration must have taken place in late- and post-glacial times. The mechanical composition of this weathered rock is shown in fig. 7. The average sorting is about 1.56 and the medians vary between 0.11 mm and 0.14 mm. The maximum grades are in the 0.125—0.25 mm class, except in the upper sample, where it is between 0.065 and 0.25 mm.

If we turn now to the soil of section I we find that in textural character it resembles that of disintegrated rocks, the similarity with the disintegrated mangerite-syenite from Holsenøy being especially notable.

The result of the mineral studies showed to a great extent a qualitative and quantitative similarity between the soil and the underlying rock. This is shown graphically in fig. 8. There is, however, one marked difference, the soil contains a considerable amount of mica, mainly a golden biotite, while the underlying rock is very poor in this mineral.

The presence of biotite in the soil might be explained by a chemical transformation of hornblende into biotite, (with an addition of potassium). It has not, however, been possible to find any evidence of this, either in the thin sections, or in the study of the grains. In the 0.065—0.25 mm fraction the hornblende content of the rock is 59.7 per cent and the overlying soil 50.3 per cent respectively. The biotite content of the rock is almost nil, while the biotite content of the overlying soil is 21.7 and 20.8 per cent respectively. It follows that if all the biotite were formed by transformation from hornblende a much smaller percentage of hornblende in the soil could be expected.

A more likely explanation of the presence of biotite in the soil is that it is derived from hidden mica-bearing zones in the underlying rock. This assumption may be supported by the fact that among the coarse grains of the soil there are also flàky mica-bearing schists.



Mineral fraction 1 - 0.25mm. Mineral fraction 0.25-0.065mm. Mineral fraction 0.065-0.01mm.

Fig. 8. Histograms showing mineral distribution in soil at Midtre Gjevilvasskamm. Section I.

A certain movement in the soil and mixing of mineral particles is to be expected. Samples collected from rock exposures surrounding the soil-covered plateau were, however, all hornblende-schist containing only small amounts of biotite.

As will be seen, the conclusion regarding the origin of the soil is not beyond dispute, but the most likely explanation does seem to be that the soil is derived from the underlying rock by weathering.

Chemical weathering.

Owing to low temperatures, chemical weathering is usually not regarded as being of any importance in the formation of alpine soils. A certain amount of chemical weathering does, however, take place, as was shown by Williams (1949), who found a concentration of CO_2 within snowbanks at high altitudes, and rock minerals in solution in melt-water from snowdrifts.

The study of heavy minerals in soil profiles has been used to a great extent in problems concerning soil maturity or stage of weathering. (O. Tamm 1916, Marshall 1940, Marshall and Haseman 1942, Humbert and Marshall 1943, Haseman and Marshall 1945, Marshall and Jeffreys 1945, R. Weil 1951 a, 1951 b, 1952, and others). The relative resistance of mineral species to the weathering processes have been discussed by a number of workers, and has been summarized by Goldich (1938): as a group the mafic minerals are less stable than the salic minerals: the Ca-rich feldspar are less stable than the alkali feldspars, and the sodic feldspars are less stable than the potash feldspars. Goldich arranged the common rock-forming minerals in a ''mineral stability series'', which is identical in arrangement with Bowen's reaction series, but which should not be interpreted as a reaction series.

Olivine — Calcic plagioclase
Augite — Calcic-alkali plagioclase
Hornblende — Alkali-calcic plagioclase
Biotite — Alkali plagioclase
Potash feldspar
Muscovite
Quartz

Dryden and Dryden (1946), comparing the heavy minerals in certain fresh rocks and their weathering products, showed the stability of the common heavy minerals as follows:

Zirkon > Tourmaline > Sillimanite > Monazite > Chloritoid > Kyanite > Hornblende > Staurolite > Garnet > Hypersthene.

We may in this connection mention that garnet is regarded by most workers as being a more stable mineral than was suggested in this work.

The two heavy minerals of common occurrence which are most resistant to weathering and which are least likely to arise through secondary processes are zirkon and tourmaline. Epidote is usually regarded as being a more stable mineral than hornblende. Hornblende, because of its wide distribution and moderate resistance to weathering forces, is a good mineral to use when comparing the severity of weathering of different horizons of a soil. (Haseman and Marshall 1945, Cady 1940).

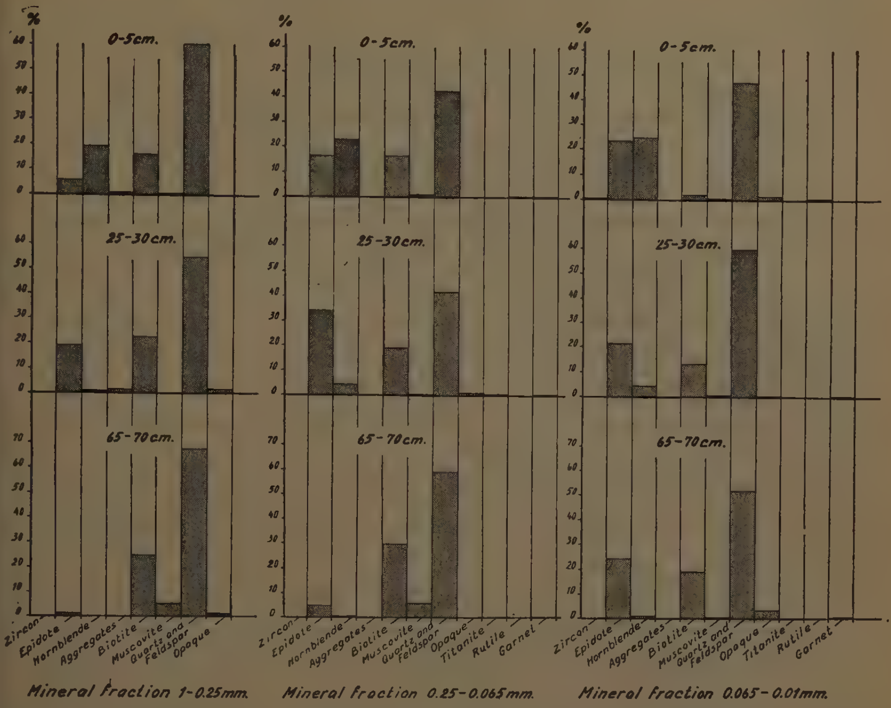
When trying to obtain a quantitative measure of the degree of weathering it is important to have a homogeneous parent material. Characteristic of differences in the degree of weathering are qualitative and quantitative similarity in the suite of resistant heavy minerals, and qualitative similarity but quantitative differences in the more easily weathered minerals.

In the soil of section I the study of the thin sections and the grains showed that a certain amount of chemical weathering has taken place. In the thin sections of the rock the hornblende grains were commonly surrounded by a light brownish alteration zone due to the oxidation of ferrous iron. Hydrated ferric oxide was also found in cracks and had stained the rims of quartz and feldspar. Brown staining zones were also observed on hornblende grains in the soil. That a certain amount of chemical weathering has taken place is also shown by the biotite, which shows some degree of bleaching.

To obtain a quantitative measure of the degree of chemical weathering of the soil is difficult. The resistant heavy minerals, such as zirkon and tourmaline, do not occur in sufficiently large amounts to be of much value, and also the parent material can not be considered as homogeneous. The relative amounts of epidote and hornblende might, however, give an indication of chemical weathering, since epidote is a more stable mineral than hornblende. From the mineral studies it will be seen that there is an increase in the epidote content of the lower sample of the soil. This is especially marked in the 0.065—0.01 mm fraction. The weight proportions between epidote and hornblende are as follows:

0.25—0.065 mm fraction.

Epidote:	hornblende	0— 5	cm	1: 22.3
»	»	5—25	»	1: 6.1
»	»	Rock	»	1: 30.0



Mineral fraction 1-0.25mm. Mineral fraction 0.25-0.065mm. Mineral fraction 0.065-0.01mm.

Fig. 9. Histograms showing mineral distribution in soil at Midtre Gjevilvasskamm. Section II.

0.065—0.01 mm fraction.

Epidote: hornblende 0— 5 cm 1: 12.7
» » 5—25 » 1: 1.55
» » Rock 1: 6.7

The values of hornblende relative to epidote in the 5—25 cm sample may indicate that some of the hornblende grains have been lost by chemical action.

As a conclusion it may be said that the chemical weathering which has taken place in section I is of a low degree. This is also shown by the very low percentage of material in the clay fraction.

Physical weathering.

Physical weathering is no doubt the most important factor in the formation of the soil, and is largely connected with the changes in temperature. The physical weathering may be due either to temperature weathering through changes in insolation and exradiation without the action of water, or to frost weathering due to freezing and thawing of water in fissures. Experiments by Griggs (1936), Taber (1943), and Blackwelder (1925, 1933) have not shown temperature changes alone to be a very important factor in the disintegration of rocks and the most important factor is doubtless the shattering of rocks by freezing and thawing of water which has entered cracks. The rocks at Midtre Gjevilvasskamm are very susceptible to frost weathering: they consist of schists or schistous gneisses and have a dip of c. 30°, so that the water will easily penetrate into the planes of weakness of the rocks. As will be seen from the temperature observations, frost cracking can be expected to be most frequent in the spring and autumn, when temperatures frequently lie around freezing point. Diurnal temperature variations in rocks have, however, been shown to be much greater than in air (I. DIRMHIRN, 1952), and are to a great extent dependent upon the clearness of the sky and on the wind conditions.

The temperature measurements by I. Dirmhirn in the Alps, at an altitude of 3050 m, for instance, showed that on a clear, calm day in September the maximum temperature on a slab of gneiss was almost 29° C, with a corresponding air temperature of 6.3° C. At night, the minimum temperature on the gneiss slab was about ÷ 4° C, with a corresponding air temperature of 1.8° C.

This means that freezing and thawing must take place very much more frequently than is shown by the air temperature. I Dirmhirn's temperature investigations also showed that the physical weathering progresses more rapidly the thinner the rock slab is.

Melt-water from snow drifts may help to maintain saturation at and close to the surface. Further down, the cycle of freezing and thawing occurs only once a year, and this limits the depth of residual soil formed by this process. The result of prolonged action of freezing and thawing is thought to be a well-sorted soil, with an increasing degree of fineness. This point is, however, not absolutely clear, and obviously fineness will also depend on the type of parent material.

In conclusion it may be said that the soil (section I) at Midtre Gjevilvasskamm is probably a residual soil, weathered *in situ,* mainly by the action of frost weathering.

SOIL SECTION II.

The origin of the soil.

From the histograms and cumulative curves (figs. 5 and 6) it will be seen that there is a very marked difference in the mechanical composition in the different soil horizons. While the top part is well sorted, with a sorting coefficient of 1.73, the sorting is less good in the middle sample, the So being 2.51, while in the bottom sample the sorting is poor, with a So of 4.76. The medians are also different, being 0.19 mm in the upper, 0.17 mm in the middle, and 0.7 mm in the bottom sample. The upper and middle samples are unimodal, with their modes in the 0.125 mm class, while the bottom sample is polymodal. It will be noted that the modes are in the same class as those in Section I.

From the mechanical composition alone it might be suggested that the upper part of the deposit section was eolian in origin. There is, however, no lamination to be seen; moreover, the grains lack rounding and pitting, and also there is a certain amount of coarse material. From tables 6, 7, 8, and the histograms (fig. 9) it will be seen that the upper sample contains a large number of mineral types and that the proportion of minerals in the three fractions show a definite relationship. All these facts seem to go to show that wind action has not been of any importance in the formation of the soil.

The deposit might also be considered as a residual soil, weathered *in situ* from the underlying rock. The bottom sample would in that case represent a residium where disintegration had not progressed very far, and the two other samples would show more advanced disintegration, with a maximum at the top. A study of the coarse part of the bottom sample ($>$ 8 mm), shows, however, that the pebbles are definitely sub-rounded and show distinct signs of wear. The pebbles consist of hornblende-bearing and biotite-bearing gneisses and a few mica-schists. The most likely supposition is that they are transported and deposited by the ice.

The mechanical composition of the bottom sample has also great resemblance to that of moraine material. The much higher percentage of clay and silt (fig. 3) in this sample than in the bottom sample of Section I may also suggest a glacial origin.

The pebbles present in the two samples nearer the surface have, through disintegration and with a few exeptions, lost the form and features which were so typically exhibited in the lower sample. But in the upper sample especially there are a few ice-worn relics, and pebbles consisting of a number of rock types foreign to the area, which seems to suggest that the material is a strongly disintegrated till.

In table 5, which shows the percentages of heavy minerals at different soil-depths, it will be seen that the two upper samples have a H.M. content which differs markedly from that of the bottom sample.

This difference is also very marked when one studies the proportion of mineral species, (tables 6, 7, 8, and fig. 9) but it will be seen that there is also a very marked difference between the two upper samples.

There is, on the whole, a qualitative similarity in the mineral suite at different depths of the soil. There is, however, a great quantitative difference: the upper sample is characterised by a fairly equal amount of hornblende and epidote, and apart from quartz-feldspar there is also some biotite. In the middle sample the hornblende content is very low, whilst the epidote content is very high; the biotite and quarts-feldspar content does not differ very much from the upper sample. In the bottom sample the hornblende has almost disappeared, the epidote content is low, except in the finest fraction, the biotite content has increased, and muscovite is also present in a greater amount than in the upper samples.

The most common rock-types found in the coarser parts of the upper sample are hornblende-schists and hornblende-bearing quarts-feldspar rocks. In the middle sample the pebbles consist mainly of a schistous quartz-feldspatic rock containing a great amount of epidote in addition to biotite and hornblende. In the bottom sample the majority of the pebbles consist of biotitebearing gneisses, and mica-schists. There seems, therefore, to be a correspondence in mineral content between the coarse pebbles and the fine parts of the sample.

The most likely conclusion is, therefore, that the deposit was originally a till, the upper part of which has undergone a certain amount of disintegration. The heterogenity with regard to the lithological composition of the till is shown in the coarse material as well as in the fine.

Weathering.

Owing to the heterogeneous character of the soil it is not possible to obtain a quantitative measure of the degree of weathering. The freshness of the minerals and the presence of an easily weathered mineral like hornblende in the upper parts of the soil, show, however, that the chemical weathering is not of any great significance.

Physical disintegration, especially that due to frost, is, without doubt, here as in section I the most powerful process of weathering.

Remarks on the glaciation of the area.

It was mentioned at the beginning of this paper that certain investigators are of the opinion that Midtre Gjevilvasskamm and other summits in Trollheimen were nunataks during the last glaciation. (N. A. SØRENSEN 1949, A. GRØNLI 1950). It has also been suggested that the soil accumulation on the southern part of Midtre Gjevilvasskamm was pre-last glaciation (Wisconsin) in age. Without entering into a full discussion of the problem of the maximum thickness of the ice sheet in this area during the last glaciation, some facts may be put forward; glacial striae with a direction NNW were found on the northern, steep slope of the mountain up to an altitude of 1450 m. It was also found that the freshness and distinctiveness of the striae gradually decreased with increasing altitude, and that the disintegration of the rocks gradually increased with the altitude. Above the highest observed striae, right up to the northern summit, the rocks had still the relic forms of roches moutonnées. The soil on the summit plateau, section II, was found, on analysis, to be suggestive of a partly disintegrated till.

From these considerations it seems safe to conclude that the whole area has, at some time, been covered by the inland ice sheet. The problem, however, is whether the area was overridden by the ice sheet during the last glaciation. In the author's opinion the facts seem to point to this. Firstly the glaciated forms of the rocks are found right up to the northern summit, in an area where local glaciation is not likely to have occurred. Secondly, the disintegration of the rocks shows a gradual increase from the foot of the mountain to the summit. If the inland ice did not extend much higher than 1450 meters, it is thought that the disintegration of the rocks would have obliterated any glacial forms from en earlier glaciation higher than this level.

With regard to the soil at the summit of Midtre Gjevilvasskamm, the age cannot be established with any degree of certainty. Knowledge of soil formation processes and rate of weathering in alpine regions is as yet very limited. It is, however, thought that if soils represented deposits of pre-Wisconsin age, their degree of weathering, both physical and chemical, would be greater. Finally, it is doubtful whether the soils, which are in a rather exposed position, would not have been removed by wind and solifluction during the long spell of time.

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Manuscript received January 15, 1953.

NOTISER

***Glossomorphites*, a new generic name for *Glossopsis* Hessland, 1949, preoccupied. (Ostracoda).**

I. HESSLAND
(Louisiana State University, Baton Rouge).

The generic name of *Glossopsis* was proposed in 1949 (HESSLAND: Lower Ordovician ostracods of the Siljan district, Sweden; Bull. Geol. Inst. Upsala, Vol. XXXIII, p. 296).

Mr. G. HENNINGSMOEN (Paleontological Museum, Oslo) has kindly called my attention to the fact that this name, viz. *Glossopsis* HESSLAND, 1949, is preoccupied.

KATHERINE J. BUSCH had previously applied *Glossopsis* to a Polychaet genus in her work on "Tubiculous Annelids of the tribes Sabellides and Serpentides from the Pacific Ocean" in Harriman Alaska Expedition, with cooperation of Washington Academy of Sciences, Vol. XII, p. 225—226, New York 1904.

Glossomorphites is now proposed as a new name for the ostracod genus *Glossopsis* HESSLAND, 1949. Like *Glossopsis* it refers to the fact that the anterior lobe (L I) is often linguiform. Type species: *Glossopsis lingua* HESSLAND, 1949.

As a consequence of this change of the generic name the name of the subfamily Glossopsiinae HENNINGSMOEN, 1953 has to be altered to Glossomorphitinae (HENNINGSMOEN: Classification of Paleozoic straight-hinged ostracods; Norsk geol. tidsskr., Vol. 31, 1953, p. 201).

NORSK GEOLOGISK TIDSSKRIFT 32

Clausthalite and Selenium-bearing Galena in Norway.

BY

Knut Heier

Selenium minerals have not previously been known for certain from any Norwegian locality, though it has been thought to exist in some mineral veins in Telemark and Setesdal, South Norway. Professor J. H. L. Vogt[1] mentions clausthalite from an unknown locality in that region. An analysis shows this mineral to be rather rich in silver, but no analytical data are given. Further investigations to confirm the identity were not carried out.

During the work to determine the exact X-ray powder patterns of the sulphide and sulphosalt minerals, a specimen from the collections in the Mineralogical Museum, Oslo, labeled arsenopyrite from Numedal, turned out to be a mineral belonging to the galena group with a spacing between those of galena and clausthalite. Another specimen from the same locality showed a spacing very close to that of clausthalite. In order to find more localities of this mineral, approximately 20 galena specimens from the Kongsberg Silver Mines were examined. One of these showed a spacing very close to the first specimen from Numedal. The exact locality was Herzog Ulrich Mine, which is situated on the Underberget and according to C. Bugge[2] belongs to the Gottes Hülfe vein. This last locality is particularly interesting as the Kongsberg Silver Mines has been examined by several geologists, and the mineral paragenesis has been thought to be very well known.

Galena and clausthalite are isomorphous and J. W. Earley[3] has proved the existence of the solid solution series PbS-PbSe by making fusions at 20 molecular per cent intervals between the end members and then calculating the cube edge from the back reflection lines on powder diagrams. He gives the following unit cell dimensions:

	Pbs	20 mol. %	40	60	80	PbSe	
a =	5.923	5.959	6.002	6.038	6.076	6.110	kx

By plotting these values on a diagram with the composition as the absissas and the cube edge values as the ordinates an almost linear curve is obtained. The diagram is reproduced in this paper as fig. 1.

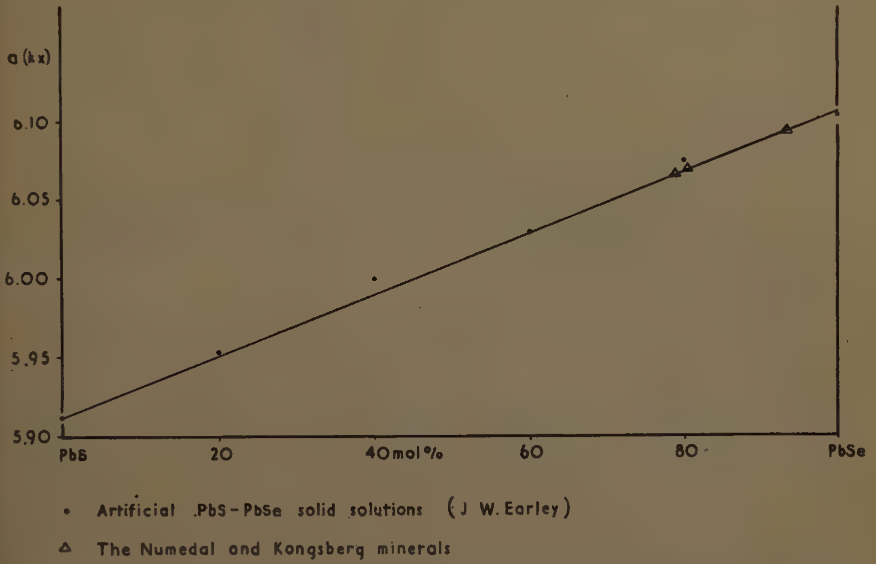

Fig. 1. Plot of cube-edge and composition of artificial PbS-PbSe solid solutions. J. W. EARLEY (3).

Calculation of the cube edge for the three Norwegian minerals produces the following result:

Mineral 1 Herzog Ulrich mine Kongsberg a = 6.070± .003 kx
2 Numedal a = 6.065± .003 kx
3 Numedal a = 6.098± .003 kx

By plotting these values on the diagram the following compositions are obtained:

Mineral 1 80 mol. % PbSe
2 79 —
3 93 —

The diagram is based on pure artificial compounds while naturally occurring minerals might contain impurities which influence the spacings. The one element which will produce the same effect as selenium is tellurium. In order to check this, mineral 3, which is the most selenium-rich, was given to Mr. S. Rutlin at the Sentralinstituttet for Industriell Forskning, Oslo, for optical spectrographic analysis, from which no tellurium was discovered. The examination gave the following result: (as will be well known selenium and sulphur do not register by this method)

Elements	Probable percentage
Pb,	much
Fe, Cu, Ca,	> 1 %
Si, V, Mo, Al, Ag, Mg,	< 1 %
Ti, Mn, Yb, Yt.	< 1/10 %

The Fe and Cu is ascribed to impurities of chalcopyrite. Mo is probably from molybdenite, and the Ca, Si, Al, Mg, Mn, and Ti are probably intermixed vein minerals. Ag most probably replaces Pb in clausthalite, while Yb and Yt replaces Ca in calcite.

Numedal is a valley running from Kongsberg in a NW direction for about 130 km. Along this valley there are three different mine districts: the North and South Vinoren which belong to the Kongsberg Silver Mines, and the molybdenite and copper mines situated farther up the valley. The question now arises as to which of these localities the minerals belong. Another sample of arsenopyrite from the Museum collected at about the same time was labeled Åslandåsen, Numedal. Åslandåsen is the name of a series of mines in North Vinoren. This together with the fact that a selenium-bearing galena was found from Herzog Ulrich Mine, Kongsberg, points towards Vinoren as the most probable locality.

However, the possible content of molybdenite as is indicated by the analysis suggests one of the Numedal molybdenum mines as the locality since this mineral is not known from Kongsberg. Arsenopyrite is also very rare at Kongsberg, having been reported by some early geologists. However, H. NEUMANN[4], who has made a special study of the minerals from Kongsberg, has never found it, and he writes that it is very probable that in the cases where it is said to have been found, arsenides of Ni and Co (and Fe) have erroneously been taken to be arsenopyrite.

An excursion to the molybdenum mines in Numedal will be carried out in the near future, and if new results are obtained, they will be published at a later date.

Together with H. NEUMANN and R. KONGSGÅRD an excursion was made to the North Vinoren mines where all the old dumps were examined. No specimens resembling those from the Museum were found, but a sample of galena from Åsland and one from Åslandåsen gave, when examined by X-rays, a powder pattern indicating approximately 10 mol. % PbSe.

Even if there is some uncertainty as to the origin of the Numedal specimens, the Herzog Ulrich locality is beyond doubt, and it can be concluded that the Kongsberg silver deposits have a certain selenium content, which had hitherto remained unnoticed.

The minerals concerned were discovered during the X-ray examinations of sulphides and sulphosalts which were carried out by

H. NEUMANN and the writer in the Mineralogical Museum, Oslo. The work was done with a self-centering X-ray camera for powder diagrams described by H. NEUMANN and J. BARSTAD[5].

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University of Oslo, Mineralogical Museum, June 1953.

NORSK GEOLOGISK FORENING

Generalforsamling torsdag 12. februar 1953 i Drammensveien 78.

Tilstede 32 medlemmer og 6 gjester.

Årsmelding for 1952.

Siden forrige generalforsamling er foreningens medlem Berging. R. Støren avgått ved døden.

5 medlemmer er meldt ut av foreningen.

I samme tidsrom er innvalgt følgende 9 nye medlemmer:

353	Gunnar Kautsky
354	Pontus Ljungren
355	Henning Sørensen
356	Lauritz Bjerrum
357	G. K. Rutherford
358	Arne Hofseth
359	Knut Heier
360	Gunnar Gillberg
361	Erling Johansen

Foreningen har nå 206 medlemmer, herav 2 korresponderende medlemmer, 95 livsvarige, og 109 årsbetalende medlemmer.

Det er avholdt 6 møter med samlet frammøte av 231 personer.

Av Norsk Geologisk Tidsskrift er bind 30 utkommet. Bind 31 og hefte 1 av bind 32 er under trykning.

Regnskapet ble godkjent.

Valg. Resultatet av valgene ble:
Formann: *J. A. Dons.*
Viseformann: *L. Størmer.*
Sekretær: *R. Selmer-Olsen.*
Redaktør: *N.-H. Kolderup* (dessuten representant for Bergens Geologiske Klubb).
Styremedlemmer:
S. Føyn, E. Kiil, N. Spjeldnæs, Th. Vogt. Den siste er representant for (Trondhjems Geologiske Klubb).

Reuschmedaljekomité:
J. A. Dons, H. Holtedahl, T. Strand.
Revisorer:
T. Gjelsvik og *T. Strand.*

*

Styremedlemmene i Berg- og Steinindustriens Gavefond er *C. Bugge, O. Holtedahl* og *G. Holmsen.*

N.G.F.'s styremedlemmer i Sulitjelmafondet er *T. Barth* og *G. Holmsen,* varamann: *L. Størmer.*

I. Th. Rosenqvist refererte hva som var gjort i den senere tid for å realisere den plan som tidligere var fremlagt for foreningen om en internasjonal metamorfosekongress i Oslo et av de nærmeste år.

Han antydet at Kirke- og Undervisningsdepartementet hadde stilt seg velvillig til spørsmålet om å yte et bidrag til foretagendet.

Det ble referert en redegjørelse fra Chr. Gleditsch om planer fremkastet av «Association pour l'étude géologique des zones profondes de l'écorce terrestre» under nittende Internasjonale Geologkongress i Alger 1952, om en ekskursjon i prekambrium i Oslofjord-området.

En ser en mulighet i en kombinasjon av de to forslag.

Det ble videre referert et forslag om et nordisk geologmøte i Göteborg januar 1954.

Reuschmedaljen ble utdelt til *Hans Holtedahl* for avhandlingen: «A Study of the Topography and the Sediments of the Continental Slope West of Møre, W. Norway». Årbok 1950, Universitetet i Bergen.

Utdrag av regnskap for 1952.

Det ordinære budsjett.

Inntekt.

Beholdning, overført fra 1951:	I kasse	kr. 644,44	
	I bank	« 14.785,81	kr. 15.430,25
Innbetalt medlemskontingent		«	1.080,00
Tilskudd fra Staten		«	1.200,00
Tilskudd fra Norges Almenvitenskapelige forskningsråd		«	6.721,22
Abonnement og salg av tidsskriftet		«	4.257,50
Renter av bankinnskudd		«	173,47
Renter av Livsvarige medlemmers fond		«	363,82
Renter av Berg- og Stenindustriens fond....................		«	407,65
			kr. 29.633,91

Utgift.

Tidsskriftet: Trykking, Bind 28 h. 2—4 (rest)	kr. 504,00	
« « 29	« 9.631,95	
Klisjeer « 29	« 2.382,22	kr. 12.518,17
Korrektur		« 125,00
Arbeidshjelp, porto og skrivesaker		« 1.298,09
Møter og ekskursjoner		« 531,15
Rabatt ved salg av tidsskriftet		« 952,69
Representasjoner og anskaffelser		« 131,05
Beholdning, overført til 1953: I kasse	kr. 261,38	
I bank	« 13.816,38	« 14.077,76
		kr. 29.633,91

1952. Livsvarige medlemmers fond.

		Inntekt	Utgift
Fondets kapital 1/1 1952:			
Obligasjoner	kr. 500,00		
Bankinnskudd	« 12.550,00		
Innbetalt kontingent i 1952	« 1.300,00		
Urørlig kapital	kr. 14.350,00		
Renter av bankinnskudd og obligasjoner		kr. 369,82	
Forvaltning av verdipapirer			kr. 6,00
Overført renter til ordinært budsjett			« 363,82
		kr. 369,82	kr. 369,82

1952. Berg- og Stenindustriens gavefond.

Fondets kapital 1/1 1952:			
Urørlig	kr. 20.000,00		
Disponible	« 5.554,00		
Renter for 1952		kr. 407,65	
Overført til ordinære budsjett			kr. 407,65
		kr. 407,65	kr. 407,65

Regnskapet revidert av T. Strand og T. Gjelsvik, Oslo, 12/1 1953.

Møte 344 torsdag 12. februar 1953, i Dr.vn. 78.

Til stede 32 medlemmer og 6 gjester.

Innvalg av følgende nye medlemmer ble godkjent:

362 Fru *Kirsten Jakhelln*, Irisvegen 18, Berg, Oslo. Foreslått av H. Major og H. Rosendahl.

363 Vitenskapelig assistent cand.real. *Bjørn Andersen.* Geologisk Institutt Blindern. Foreslått av O. Holtedahl og I. Oftedal.

364 Professor *Henry V. Hove.* School of Geology. Louisiana State. University Baton Rouge. Foreslått av G. Henningsmoen og N. Spjeldnæs.

365 Professor *Grover E. Murray.* School of Geology. Louisiana State University Baton Rouge. Foreslått av G. Henningsmoen og N. Spjeldnæs.

366 Professor *Kenneth E. Caster.* Departement of Geology and Geography. University of Cincinnati, Cincinnati 21, Ohio. Foreslått av G. Henningsmoen og N. Spjeldnæs.

367 Dr. *Edmund Nosow.* Box 282, Kentucky Geological Survey, Hopkinsville, Kentucky. Foreslått av G. Henningsmoen og N. Spjeldnæs.

Foredrag av *H. Holtedahl* (Bergen): *«Noen sub-marin-geologiske undersøkelser utenfor Norges vestkyst».*

Med støtte av Det Almenvitenskapelige Forskningsråd ble der i 1950 og 1951 foretatt to tokter med forskningsfartøyet «Armauer Hansen», tilhørende Universitetet i Bergen, med den oppgave å drive sub-marin geologiske undersøkelser i visse strøk utenfor Norges vestkyst. Det første års undersøkelser ble foretatt på et relativt lite felt på Storegga utenfor Mørekysten, hvor der ble tatt en rekke ekkosnitt utover kontinental-skråningen samt innsamlet en rekke bunnprøver fra forskjellig dyp. På toktet i 1951 ble undersøkelsene først lagt til et område utenfor Sognesjøen og videre til kontinental-hyllen og kontinental-skråningen utenfor Mørekysten. I strøket utenfor Mørekysten ble der tatt en rekke ekkosnitt på tvers av den markerte dyprennen som skiller skjærgårdsområdet fra det utenforliggende bankeområdet mellom Breisunddjupet og Buagrunnen, og videre en rekke snitt utover kontinental-hyllen. En av hovedoppgavene ved undersøkelsene var å klargjøre bunntopografien på kontinental-skråningen, og i et område mellom 62° 40’ N og 63° 50’ N ble der tatt en rekke ekkosnitt såvel utover som langsmed skråningen.

Bunnprøver ble tatt med bunnskrape og sylinderprøvetager både utenfor Sognesjøen og Mørekysten. I alt ble tatt 47 prøver, hvorav 12 bunnpropper på opptil 80 cm’s lengde.

Utenfor Sognesjøen fikk man ved ekkosnittene et godt bilde av den bortimot 500 meter dype, N-S gående renne som finnes rett utenfor kysten i Norske Rennens østlige del. Videre ble det konstatert at den undersjøiske N-S gående rygg øst for denne renne til dels har meget bratte sider og består av fast fjell. Ved skraping fikk man opp friske, istykkerbrutte fliser av fyllitt som antaes å stå som fast fjell i ryggen. Dette viser at de devonske bergarter i Solund-området også helt i vest er begrenset av kambro-silur-bergarter.

Ekkosnittene nord for Breisunddjupet gir et tydelig bilde av den markerte renne mellom skjærgårdsområdet og bankeområdet. Snittene viser hvordan havbunnen øst for rennen består av fjell med et meget uregelmessig relieff, mens den i vest har en meget jevn overflate med sedimenter. Kontinentalskråningen utenfor Mørekysten har en meget uregelmessig topografi. I hovedtrekkene er kontinentalhyllen innskåret av to store innbuktninger, atskilt ved et framstikkende nes. Den nordlige bukt er ca. 35 km bred og den sydlige ca. 15 km. Mens kontinentalskråningen både nord og syd for de to innbuktninger er ganske slak med en relativt jevn overflate i den øvre del, er buktene begrenset av meget bratte skråninger med høyder på opptil et par hundre meter. I den nordlige del av området har buktenes sider en markert Ø-V-lig retning, mens den i syd er V-NV-lig.

Profilene som ble tatt utover skråningen på det undersjøiske nes mellom de to innbuktninger viser en tydelig trappetrinnsformig topografi. Fra kanten på kontinentalhyllen, som ligger på ca. 170 m's dyp, går der en bratt skråning, med en gradient på 16—20° ned til ca. 250 m. Derfra skråner bunnen svakt utover til ca. 600 m hvorfra igjen en bratt skråning fører ned til ca. 800 m. Hellingen av bunnen er videre ganske svak utover til ca. 900 m, hvoretter en bratt skråning igjen kan følges ned til ca. 1050 m hvoretter hellingen igjen blir svak. De bratteste skråninger kan ha gradienter på opptil 40°. Det viser seg at den mellomste brattskråningen, som på det undersjøiske nes ligger mellom 550 m og 700 m, også finnes igjen på den ellers slake og jevne skråning nord og syd for de nevnte bukter. I nord finnes skråningen mellom ca. 600 m og 800 m, mens den i syd forekommer mellom 600 m og 750 m.

Resultatet av de morfologiske undersøkelser viser tydelig at kontinentalskråningen i dette område ikke er en akkumulasjonsterrasse, men består av fast fjell, og at hovedlinjene i topografien er strukturelt eller tektonisk betinget og har nær tilknytning til de innenforliggende landområder. Den trappetrinnsformige topografi kan forklares ved at man her ute på skråningen har en serie forkastninger som kan settes i forbindelse med den tertiære orogenese som resulterte i hevningen av den Skandinaviske landmasse.

Bunnprøver ble tatt på kontinentalskråningen, blant annet på de bratteste partier, for om mulig å få opp fast fjell. I et skrapetrekk fra den øvre skrent på Storegga var der spesielt meget skarpkantet stein, som vesentlig besto av granittmateriale, særlig en grovkornet, rød granitt. Det antaes som en mulighet at dette materiale kan representere berggrunnen i dette strøk.

Sedimentene på kontinentalskråningen blir finere med økende dybde samtidig som sorteringen blir dårligere. Grus og sand forekommer ned til en dybde av ca. 600 m., hvoretter innholdet av finere materiale, inklusiv leir, øker. Bunnproppene som ble tatt ned til ca. 1000 m's dyp, viser ingen lagning eller gradert sortering, og sedimentene antaes å være dannet ved sedimentasjon fra isfjell. Flint, kritt og Oslo bergarter er funnet i overflate-lag ned til dyp på 830 m, og viser at det ikke har foregått noen sedimentasjon ned til dette dyp etter at isfjellmaterialet ble avsatt. Løsavsetningene, både på kontinentalhyllen og på skråningen er av glacial opprinnelse. Tilstedeværelsen av grove, godt sorterte sedimenter ned til ca. 600 m, og finere, usorterte sedimenter på større dyp, skyldes utvilsomt strømforholdene, med utvasking av løsavsetningene på den øverste del av skråningen og på hylleområdet.

Ved siden av de sub-marine undersøkelser ble der også foretatt kvartærgeologiske studier i de ytre kyststrøk på Møre for å få holdepunkter om hvor langt utover kontinentalhyllen innlandsisen har gått under siste istid. Innlandsisens bevegelsesretning ble funnet å være N-V-lig til nordlig i området mellom Stadt og Ålesund. På Nerlandsøy ble de høyeste isskuringsstriper, her med N-V-lig retning, funnet i en høyde av 230 m o. h., og flyttblokker ble funnet rett under toppen i ca. 400 meters høyde. Dette viser at innlandsisen under siste istid har hatt en minimumstykkelse på 400 m i disse ytre kyststrøk, og at den med en gradient på 0.5/100 skulle ha rukket ut til ytterkanten av kontinentalhyllen.

I ordskiftet etter foredraget deltok: H. Rosendahl, W. Werenskiold, K. Strøm, T. Gjelsvik og foredragsholderen.

MEDLEMSLISTE.

à jour april 1953.
[1]: korresponderende medlem.
*: livsvarig medlem.
Tallet i parentes er innvalgsåret.
(S): stifter (18. februar 1905).

Adamson, Olge J., dr. Cort Adelers gate 12, vær. 710, Oslo. (1945).
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*Balk, Robert, professor. Box 162, Campus Station, Cororro, N. M., U. S. A. (1933).
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*Bjørlykke, Harald, dr. Norsk Bergverk, Christian August's gate 15, Oslo. (1923).
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*Broch, Olaf Anton, professor. Dept. of Mineralogy, University of Punjab, Lahore, Pakistan. (1920).
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*Braastad, Johan, dr. Postboks 808, Oslo. (1913).

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Wolff, Francis, lektor. Idrettsveien 1, Bergen. (1945).
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2	korresponderende medlemmer.
101	livsvarige medlemmer.
109	årsbetalende medlemmer.
212	medlemmer.